



Evolution paléogéographique et paléotopographique du Tian Shan Chinois au Mésozoïque

Gloria Heilbronn

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présentée par

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OSUR (Observatoire des Sciences de l'Univers) – UMR 6118
U.F.R. Structures et Propriétés de la Matière

**Evolution
paléogéographique et
paléotopographique
du Tian Shan Chinois
au Mésozoïque**

**Thèse soutenue à Rennes
le 28 mars 2014**

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RESUME

Le Tian Shan est une chaîne intracontinentale d'Asie Centrale, dont la structure lithosphérique résulte de l'accrétion de divers blocs au cours du Paléozoïque. Son histoire tectonique récente est marquée par sa réactivation au Tertiaire, liée à la collision entre l'Inde et l'Asie. L'histoire topographique et tectonique de la chaîne entre les deux orogénèses majeures (Paléozoïque terminal et Cénozoïque) reste peu contrainte. Cette thèse associe deux approches, la sédimentologie et la thermochronologie basse température, dans le but de déterminer de façon qualitative l'évolution de la topographie au cours du Mésozoïque.

Le démantèlement des reliefs associés à la chaîne tardi-paléozoïque s'achève au Trias supérieur. Il est suivi d'une activité tectonique très faible au Jurassique, majoritairement transtensive et caractérisée par des taux d'exhumation très faibles dans le Tian Shan. Une surface majeure de pénéplanation se développe alors en Asie Centrale. A la limite Jurassique – Crétacé, la mise en place de cônes alluviaux indique une réactivation de la chaîne, qui n'est néanmoins pas suffisante pour être enregistrée par la thermochronologie. Cette période est caractérisée par un régime généralement extensif en Asie Centrale (jusqu'au Bassin Caspien qui s'ouvre à l'est), et précède la phase d'exhumation lente, qui suit au Crétacé inférieur. La chaîne est progressivement réactivée à partir de 100 Ma et pendant le Crétacé supérieur, ce qui pourrait correspondre à un effet retardé de la collision du Bloc de Lhassa (140 - 120 Ma). Vers 65 - 60 Ma, une phase d'exhumation rapide atteste d'une réactivation plus intense et localisée le long des principales failles. Elle est sans doute liée aux collisions de blocs le long de la marge sud-ouest de l'Asie (e.g. Bloc du Kohistan, arc du Dras, Bloc Afghan). Dans la région du Tian Shan, l'activité tectonique semble totalement s'arrêter au Paléocène permettant le développement d'un niveau majeur de calcrêtes, avant la nouvelle réactivation au Néogène.

Par conséquent, l'association des données de thermochronologie sur le socle avec la reconstruction des milieux de dépôt dans les différents bassins, montre que la paléo-chaîne du Tian Shan s'aplanit durant le Mésozoïque. Des événements tectoniques de faible envergure ont lieu en Asie Centrale, induits par les principaux mouvements géodynamiques le long des bordures entourant l'Asie. Pourtant leur enregistrement est incomplet et seule la combinaison des deux approches étudiées permet de reconstituer l'évolution topographique et paléogéographique du Tian Shan.

ABSTRACT

The Tian Shan is an intracontinental range located in Central Asia. The structure of the range formed during the Paleozoic through the accretion of several blocks. Recently the range has been reactivated due to far-field effects of the collision between India and Asia. The topographic and tectonic evolution of the range in-between these two major relief-building phases (Late Paleozoic and Cenozoic) is still poorly understood. Two different approaches are combined in this work, in order to determine the Mesozoic topographic evolution of the area, in a qualitative way.

The Late-Paleozoic range has been progressively eroded until the Upper Triassic/Lower Jurassic. Tectonic activity was relatively quiet during the Jurassic characterised by low exhumation rates. We suggest that the tectonic regime was dominated by transtension in the Tian Shan area. This period is linked to a regional peneplanation in Central Asia. At the Jurassic-Cretaceous boundary, the occurrence of alluvial fan deposits shows a reactivation of the range, though not strong enough to be recorded by low-temperature thermochronology. This period is characterised by an overall extensional tectonic regime all over Central Asia, with the opening of the Caspian Basin further west. It precedes the phase of very slow exhumation that occurs during the Lower Cretaceous. From 100 Ma and during the Upper Cretaceous, the range is progressively reactivated. We suggest that this correspond to a delayed answer of the Lhasa bloc collision (140 - 120 Ma). Around 65 - 60 Ma, a new phase of rapid exhumation attests of a stronger reactivation, localised along the major faults. This is contemporary of bloc collisions along the south-west margin of Asia, such as the Kohistan Block, the Dras arc or the Afghan Block. In the Tian Shan area, the development of calcrete features in the Paleocene suggests the end of tectonic activity, before the new reactivation in the Neogene.

The combination of low temperature thermochronology on the basement rocks and facies sedimentology in the various basins indicate that while during the Mesozoic, the Palaeo-Tian Shan topography generally flattens, some small-scale tectonic events driven by far-field effects of major geodynamic processes around the edges of Asia did occur. However, those tectonic movements did not induce enough exhumation to be recorded by low temperature thermochronometers. Only the sediment record allows their detection and detailed description.

läufsch umesunscht
eme draeum nooch
fàngsch ne net
in
läufsch àwer garn
dann
solàng wid läufsch
gets iërix
e draeum

Courir après un rêve
Et ne jamais l'attraper
Mais courir quand même
Car aussi longtemps qu'on court
Il y a quelque part
Un rêve

André Weckmann, *ixidigar* (Eds. bf)

A tous ceux rencontrés sur mon chemin, qui m'ont indiqué une direction, parfois un détour et parfois
une impasse...

... car cela fait partie de la vie et m'a aidé à grandir, à chaque fois un peu plus.

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... et vous obtiendrez ceci.

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AVANT-PROPOS

Cette thèse a été réalisée en trois ans et demi au Laboratoire de Géosciences Rennes et a été encadrée par Marc Jolivet (directeur), Cécile Robin, Laurie Barrier (IPG Paris) et Sylvie Bourquin. Trois campagnes de terrain ont eu lieu en Chine (10 semaines au total) avec la collaboration de Yingying JIA et du Professeur Bihong FU (Laboratory of Lithospheric Evolution, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing - China).

L'étude de géochimie isotopique (cf. Chapitre 4) a été réalisée avec Philippe Boulvais et grâce à l'aide d'Emilie Marchand, qui a acquis les données lors de deux courts stages (en Licence 2 et Licence 3).

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INTRODUCTION & PROBLEMATIQUE

i) Reconstruire des reliefs et paléogéographies, pourquoi et comment ?

Dans le but de reconstruire l'évolution d'un paysage sur une longue période de temps, il est indispensable d'évaluer la forme de la surface, i.e. la topographie, ainsi que la position des reliefs et des bassins, i.e. la paléogéographie. La paléogéographie synthétise non seulement les connaissances disponibles sur l'activité tectonique d'une région, mais aussi la nature des zones étudiées (e.g. bassin continental ou domaine marin pour les zones de dépôt ; e.g. chaîne de montagnes ou surface plane pour les zones de non-dépôt). Elle débouche sur la production de cartes paléogéographiques, indispensables à la reconstruction de l'histoire géologique d'une zone et à son interprétation.

En domaine continental, l'évolution du paysage peut s'enregistrer de différentes manières, soit dans les roches formant le socle (d'une chaîne de montagnes par exemple), soit dans les bassins sédimentaires. Il existe donc deux principales approches pour reconstituer l'évolution d'une région (e.g. Reinert, 2007). La première se concentre sur l'histoire thermique enregistrée par le socle et qui indique les principales phases d'exhumation. La seconde approche a pour but d'étudier les dépôts sédimentaires, qui eux enregistrent la construction et l'érosion des paléo-reliefs. Dans les deux cas, ces approches apportent de nouveaux résultats sur l'histoire géodynamique de la région.

L'histoire thermique est classiquement étudiée grâce à des méthodes géochronologiques, telles que les analyses traces de fission et l'(U-Th)/He. Elles livrent des taux d'exhumation du socle, et des modèles statistiques de l'évolution thermique de l'échantillon peuvent être déduits. Ceci permet de reconstruire les principales phases tectoniques et leur intensité relative. Cette étude propose donc une reconstitution qualitative de la topographie, car la résolution verticale reste limitée dans l'utilisation de ces méthodes.

L'étude des sédiments déposés dans les bassins adjacents à la chaîne de montagnes permet la reconstruction de l'évolution des milieux de dépôt. Ceux-ci reflètent l'érosion de la chaîne en amont, qui représente la source des sédiments érodés, puis déposés. Les milieux de dépôt varient en fonction de la distance à la chaîne et du régime tectonique, du niveau de base et du climat. Associés à des données complémentaires (telles que paléocourants, analyses pétrographiques, etc.), ils peuvent indiquer la relative position par rapport à la chaîne et l'ampleur du relief érodé.

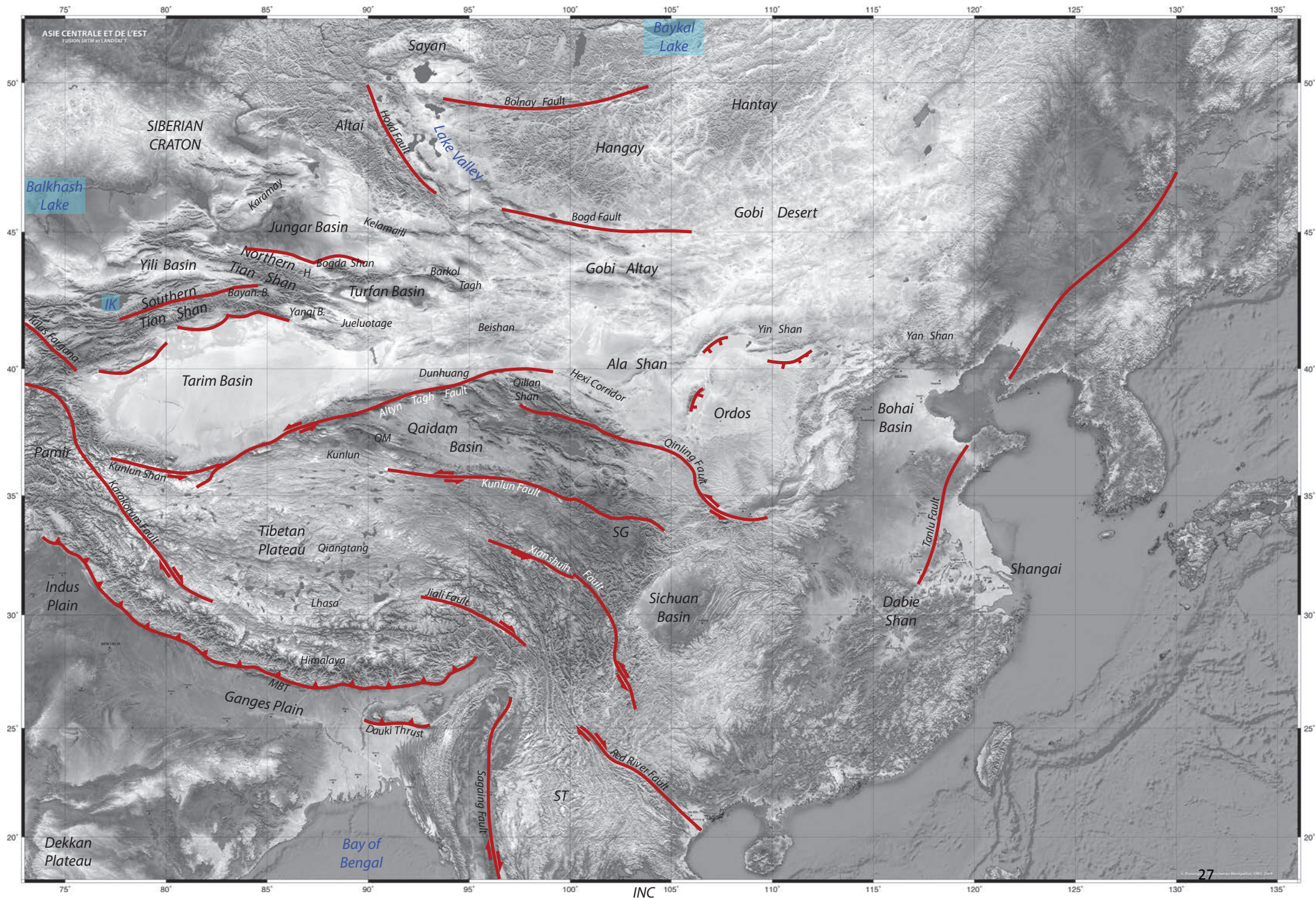
Dans le but de reconstituer au mieux l'évolution d'une chaîne de montagnes, cette thèse propose de combiner les deux approches citées, afin de disposer de meilleures limites pour la reconstitution de la paléo-topographie et paléo-géographie.

ii) Pourquoi le Tian Shan ?

L'Asie Centrale est un puzzle formé par l'accrétion de nombreux blocs au cours du Paléozoïque et Mésozoïque (Fig. i). Pour reconstituer son évolution au cours du temps, il est nécessaire de bien comprendre les principales phases d'activité tectonique (Fig. ii). Les accrétions de blocs qui ont donné lieu à la formation de l'Asie actuelle ont eu des répercussions à très grandes échelles de temps et d'espace. Afin de mieux comprendre ces événements tectoniques majeurs, les chaînes de montagnes d'Asie Centrale constituent un laboratoire naturel hors-normes. Ainsi, la chaîne du Tian Shan, située entre le Bassin du Junggar au nord et le Bassin du Tarim (au sud) a fait l'objet de nombreuses études. La formation de la chaîne au Paléozoïque a été largement étudiée (e.g. Windley et al., 1990 ; Laurent-Charvet, 2001 ; Wang et al., 2007a,b, 2009; Charvet et al., 2011 ; Xiao et al., 2013), ainsi que la réactivation cénozoïque qui a permis la formation d'une chaîne, dont l'altitude moyenne s'élève aujourd'hui à 4000 m (e.g. Avouac et al., 1993 ; Allen et al., 1999 ; Charreau et al., 2008). L'étude de la réactivation cénozoïque de la chaîne montre l'importance des structures paléozoïques et mésozoïques héritées (e.g. Allen et al., 1993a ; Allen & Vincent, 1997 ; Buslov et al., 2008 ; Jolivet et al., 2010). Il est donc indispensable de bien contraindre l'évolution de ces périodes. Notons que cette thèse porte sur le Tian Shan est, située en Chine, et que l'évolution de la chaîne sera discutée dans son ensemble dans la discussion générale de cette thèse.

Figure i : Carte de l'Asie Centrale et Orientale (Fond de carte fusion SRTM et Landsat 7 ; projection de Mercator, réalisée par S. Dominguez, 2009) indiquant les principaux bassins et chaînes de montagnes, ainsi que les principales failles. Afin de ne pas surcharger la carte, les failles n'ont pas été dessinées dans la région du Tian Shan. Le lecteur est invité à consulter les cartes plus détaillées qui suivent dans cette thèse.

SG: Songpan Gârze ; ST = Shan Tai ; INC = Indochine ; IK: Issyk-Kul Lake ; H: Houxia Valley ; Bayan. B = Bayanbulak Basin ; QM = Qimen Tagh



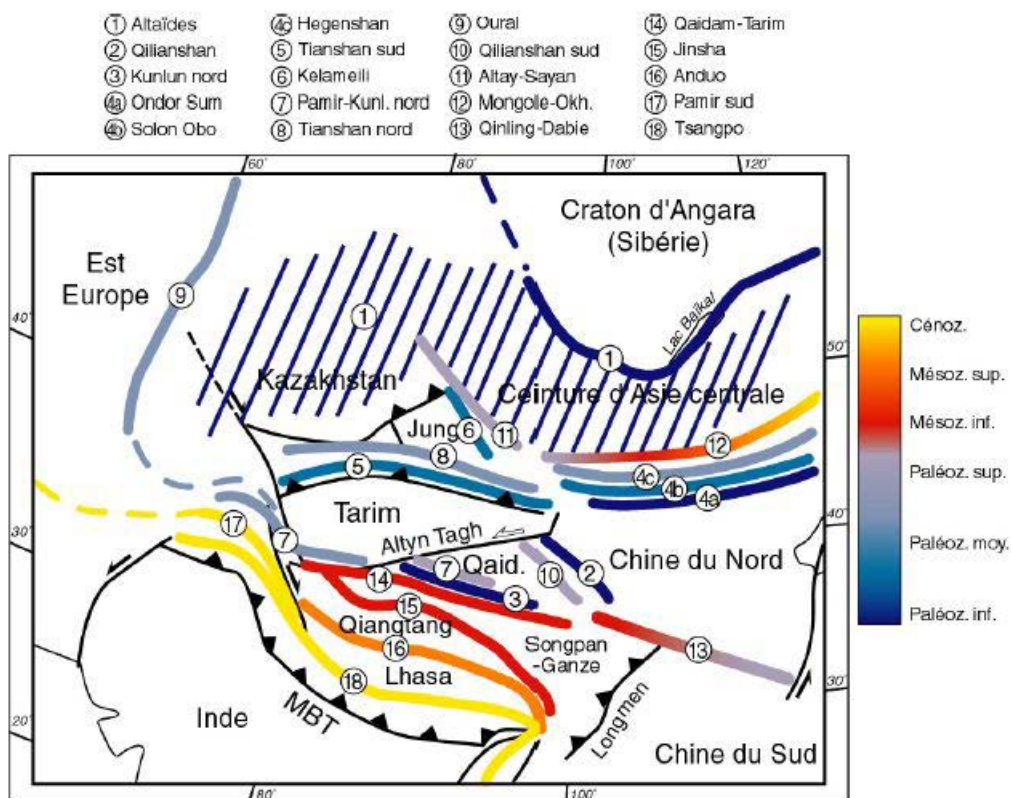


Figure ii : Synthèse des principales sutures et leur âge de formation (Laurent-Charvet, 2001 et références associées).

Entre les deux principales phases d'activité tectonique (paléozoïque et cénozoïque), l'époque mésozoïque a certes suscité de l'intérêt, mais elle est moins bien connue. Les dépôts mésozoïques de plusieurs bassins entourant la chaîne (e.g. S-Junggar, N-Tarim, Turfan) ont fait l'objet d'études sédimentologiques (e.g. Graham et al., 1990 ; Hendrix et al., 1992 ; Shao et al., 1999 ; Hendrix et al., 2000 ; Eberth et al., 2001 ; Vincent et Allen, 2001 ; Greene et al., 2001 ; Shao et al., 2003 ; Greene et al., 2005 ; Bian et al., 2010). La structure mésozoïque des bassins a également été étudiée (e.g. Allen et al., 1991 ; Allen et al., 1993a ; Allen et al., 1995 ; Allen & Vincent, 1997), tout comme l'exhumation de la chaîne (e.g. Dumitru et al., 2001 ; Sobel et al., 2006a,b ; De Grave et al., 2007 ; Wang et al., 2008 ; Wang et al., 2009 ; Glorie et al., 2010 ; Jolivet et al., 2010 ; De Grave et al., 2013 ; Macaulay et al., 2013). La synthèse de ces travaux montre que les conclusions sur l'évolution mésozoïque de la chaîne ne sont pas toutes concordantes. Un apparent paradoxe oppose une intense activité tectonique déduite de l'observation des dépôts sédimentaires au Jurassique supérieur par exemple (e.g. Hendrix et al., 1992) et un calme tectonique enregistré dans les données de thermochronologie basse-température (e.g. Jolivet et al., 2010).

Pour élucider ce problème et les différences d'enregistrement entre les deux approches citées précédemment, cette thèse propose un état de l'art non-exhaustif des événements majeurs affectant l'Asie durant le Mésozoïque (Chapitre 1), suivi d'une étude des sédiments mésozoïques observés dans plusieurs bassins autour de la chaîne actuelle (Chapitre 2). Le Chapitre 3 présente une synthèse des données traces de fission sur apatite à l'échelle de l'ensemble du Tian Shan, qui permettra de discuter les phases d'exhumation observées. Dans le but de mieux contraindre la topographie mésozoïque, une étude d'isotopie stable a été réalisée sur l'ensemble des marqueurs pédogénétiques échantillonnés (tels que nodules carbonatés de plaine alluviale, mais également des calcrètes). Les principaux résultats sont résumés dans le Chapitre 4. Enfin, les résultats issus de la combinaison des deux approches mentionnées plus haut (bassin et chaîne) permet la reconstitution de l'évolution de la paléogéographie mésozoïque du Tian Shan chinois et sera discutée par rapport au contexte tectonique générale de l'Asie Centrale (Chapitre 5).

CHAPITRE 1 - ETAT DE L'ART

1.1) Histoire géodynamique de l'Asie

La géodynamique du Mésozoïque est marquée par l'apogée, puis le démantèlement de la Pangée. A partir du Permien, ce supercontinent rassemble la quasi-totalité des terres émergées. Seul le bloc Chine du Sud, une partie de l'Indochine et les différents blocs cimmériens sont séparés du Gondwana (e.g. Qiangtang, Sibumasu, Ouest Sumatra, Est Malaya) (Fig. 1.1.1) (e.g. Metcalfe, 2006, 2013 ; Roger et al., 2010). La Pangée est alors entourée par un vaste océan, la Panthalassa (ou Paléo-Pacifique), dont le golfe formé à l'est du méga-continent se nomme la Paléotéthys. Ce dernier borde les blocs continentaux qui formeront l'Asie actuelle (Fig. 1.1.1). Au cours du Trias, les blocs continentaux cités précédemment sont amalgamés au sud de l'Asie lors de l'orogénèse cimmérienne au sens large (e.g. Mattauer et al., 1992 ; Roger et al., 2010, 2011 ; Metcalfe, 2006, 2013). Le supercontinent ne peut rester stable en raison de la convection qui a lieu dans le manteau et qui entraîne la mobilité des plaques. Son morcèlement commence au Trias supérieur. A l'échelle globale, cette période est marquée par des phases de rifting consécutives, notamment le rift de l'Atlantique nord qui sépare les deux nouveaux supercontinents : le Gondwana au sud et la Laurasia au nord (Fig. 1.1.2) (Metcalfe, 2013). En Asie, le régime tectonique général reste compressif, comme l'atteste la fermeture de l'océan Paléotéthys (Fig. 1.1.2). Le Jurassique et le Crétacé y sont marqués par deux événements majeurs : la fermeture de l'océan Mongol-Okhotsk au nord et les multiples amalgamations de blocs (e.g. bloc de Lhasa, arc du Kohistan-Dras) sur les marges sud (e.g. Golonka, 2004 ; Burg, 2011). Au Crétacé, l'Inde se détache de l'Australie et de l'Afrique et entame sa dérive vers le nord, avant d'entrer en collision avec l'Asie. L'âge de la collision fait l'objet de nombreuses discussions et est généralement estimé soit à 65 Ma (e.g. Yin and Harrison, 2000), soit vers 50 Ma (e.g. Rowley, 1996, 1998 ; Najman et al., 2001). Cet événement influence considérablement la tectonique, la topographie et le climat en Asie (e.g. Molnar and Tapponnier, 1975 ; Tapponnier et al., 2001 ; Yin, 2010 ; Searle et al., 2011). Il apparaît clairement que certaines structures tectoniques majeures sont héritées et conditionnent fortement la localisation des déformations plus récentes (cénozoïques et actuelles, e.g. Allen et al., 1993a, 1997 ; Harrowfield & Wilson et al., 2005 ; Yin, 2010 ; Buslov et al., 2008 ; Roger et al., 2010, 2011 ; Jolivet et al., 2010, 2013a).

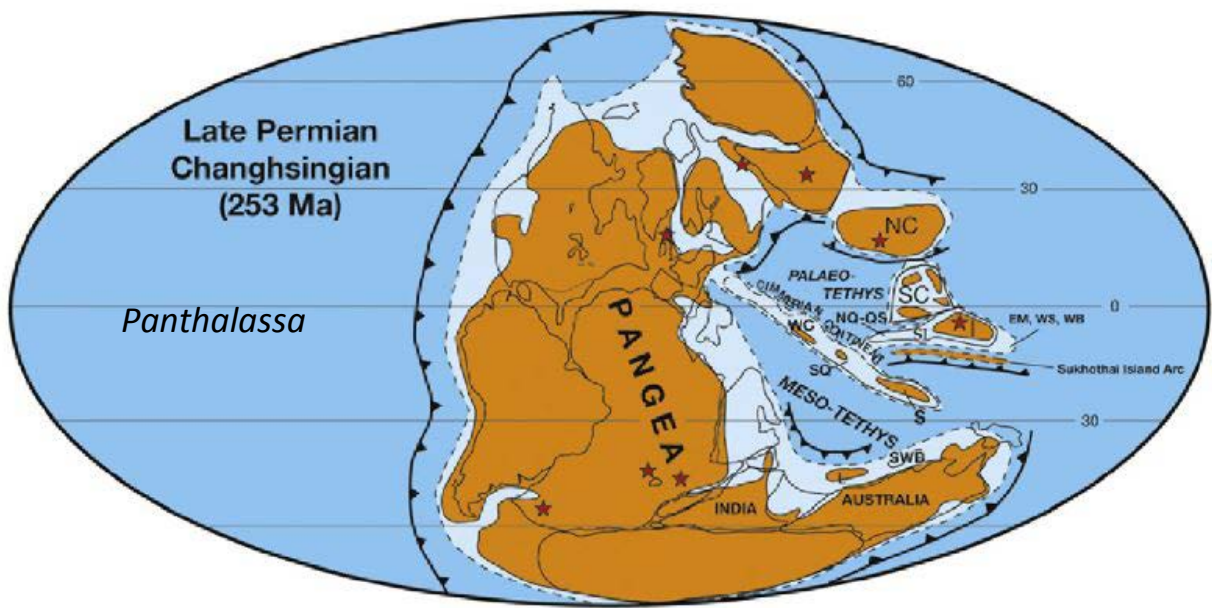


Figure 1.1.1 : Position des continents à la fin du Permien, peu avant l'apogée de la formation du supercontinent Pangée.
 SC = Chine du Sud ; T = Tarim ; I = Indochine ; EM = Malaya Est ; WS = Sumatra Ouest ; NC = Chine du Nord ; SI = Simao ; S = Sibumasu ; WB = Burma Ouest ; SQ = Qiangtang Sud ; NQ-QS = Qiangtang Nord – Qamdao–Simao ; L = Lhasa ; SWB = Borneo Sud Ouest ; et WC = Continent Ouest-cimmérien (Metcalf, 2013).

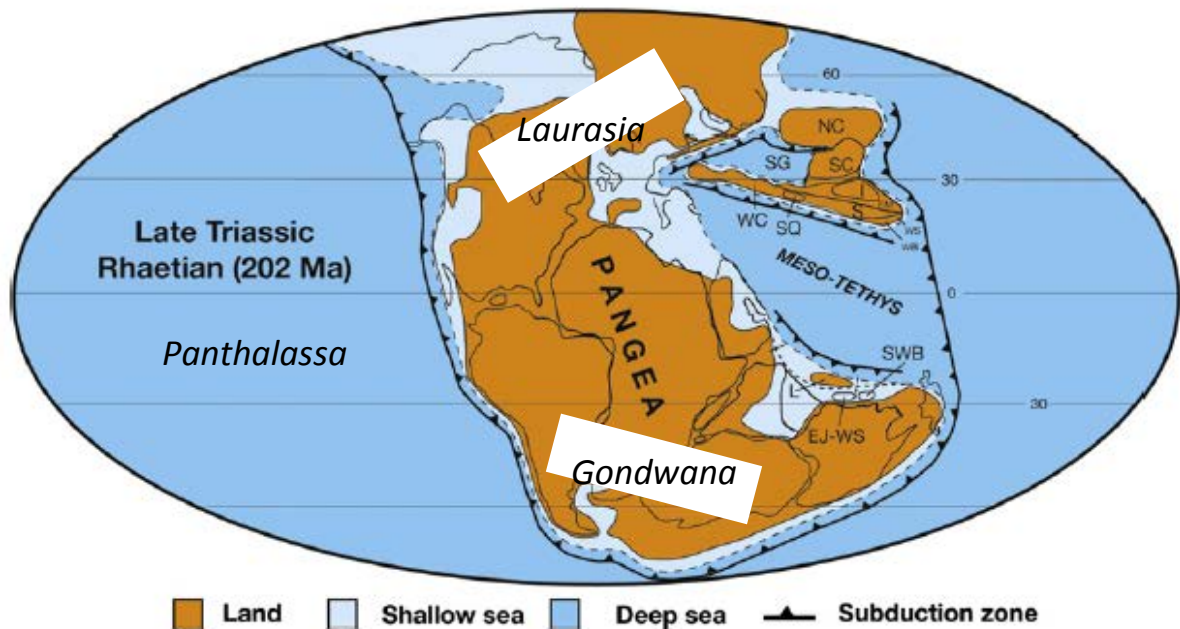


Figure 1.1.2 : Position des continents à la fin du Trias, au début du démantèlement du supercontinent Pangée, qui sera ensuite divisé en deux continents (Laurasia et Gondwana), toujours bordés par la Panthalassa et la Téthys. SC = Chine du Sud ; T = Tarim ; I = Indochine ; EM = Malaya Est ; WS = Sumatra Ouest ; NC = Chine du Nord ; SI = Simao ; S = Sibumasu ; WB = Burma Ouest ; SQ = Qiangtang Sud ; NQ-QS = Qiangtang Nord – Qamdao–Simao ; L = Lhasa ; SWB = Borneo Sud Ouest ; and WC = Continent Ouest-cimmérien (Metcalf, 2013).

De la même manière, l'histoire tectonique mésozoïque de l'Asie est influencée par l'histoire paléozoïque qui la précède (e.g. Jolivet et al., 2001, 2010 ; Sobel et al., 2001 ; De Grave et al., 2007 ; Carter & Clift, 2008 ; Glorie et al., 2010 ; Roger et al., 2010, 2014). Cette dernière est largement conservée dans la Ceinture Orogénique d'Asie Centrale (CAOB, *Central Asian Orogenic Belt*) (Fig. 1.1.3), formée à la suite de l'accrétion de divers ensembles structuraux (arcs insulaires, bassins d'avant- et d'arrière-arc, prismes d'accrétion, marges passives, microcontinents, ainsi que des unités ophiolitiques) principalement le long des marges sud et est du craton protérozoïque stable de la Sibérie. L'accrétion a débuté au début du Néo-Protérozoïque et a pris fin au Permien supérieur – Trias inférieur (e.g. Coleman, 1989 ; Zonenshain et al., 1990 ; Sengör & Natal'in, 1996 ; Windley et al., 2007). Aujourd'hui, ils forment le cœur du continent Asiatique. La CAOBB comprend l'ensemble des blocs amalgamés, situés entre le craton d'Europe de l'est au nord-ouest (bordé par la chaîne de l'Oural, indiquée par les « Uralides » sur la Fig. 1.1.3), le craton Sibérien au nord-est (bordé par la chaîne des Baïkalides) et les cratons du Tarim et de la Chine du Nord, au sud (Fig. 1.1.3, Windley et al., 2007). La CAOBB s'étend donc d'ouest en est, depuis la chaîne de l'Oural jusqu'à la marge Pacifique, et du nord au sud depuis le craton Sibérien à la chaîne du Tian Shan et la suture de Solonker (Fig. 1.1.3). La chaîne du Tian Shan constitue donc la branche principale de la partie sud-ouest de la CAOBB.

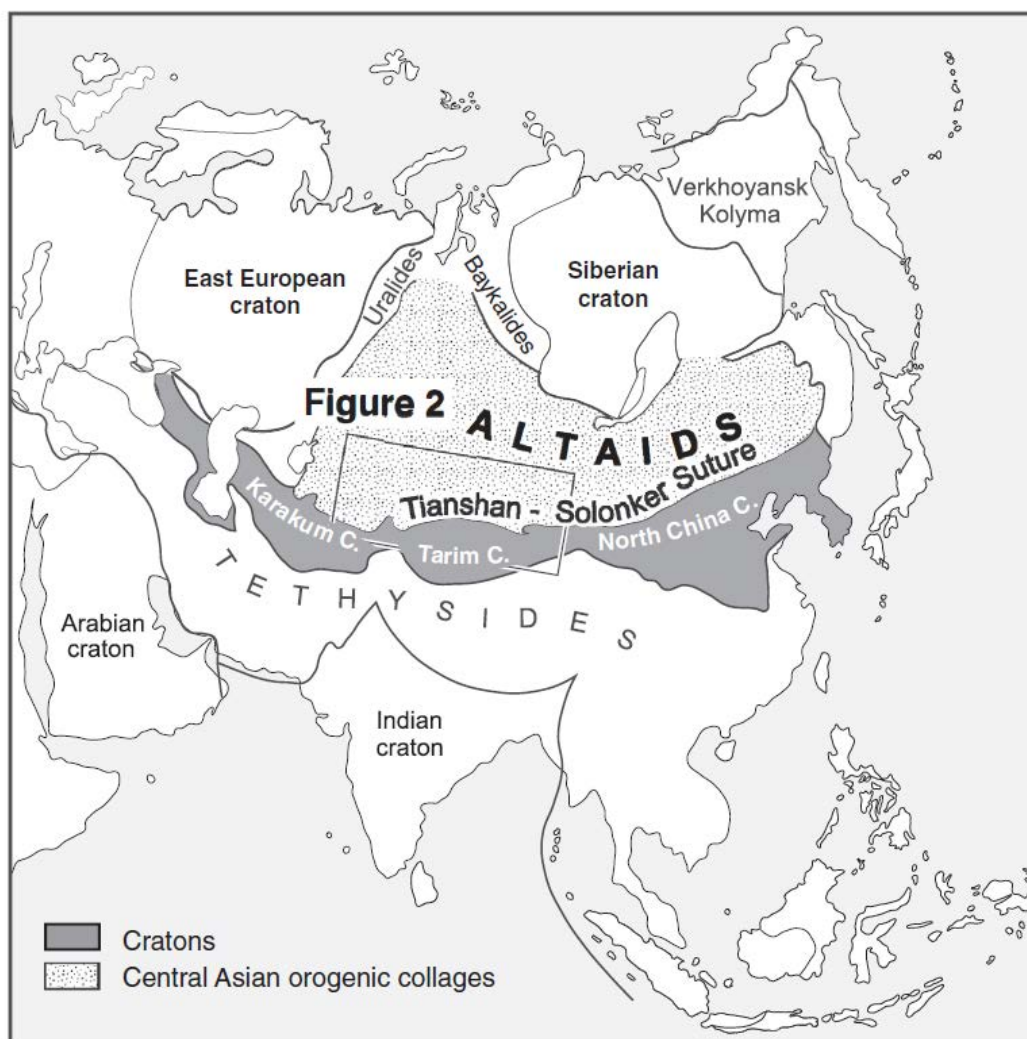


Figure 1.1.3 : Carte tectonique simplifiée de l'Asie, illustrant l'étendue de la Ceinture Orogénique d'Asie Centrale (CAOB), également appelée « Altai » (Xiao et al., 2013).

Le début et la fin du Mésozoïque sont des temps forts dans l'histoire de l'Asie, car ils sont marqués par des collisions suivies d'orogénèses de grande ampleur. Au sud, les deux collisions majeures sont la collision entre le bloc de Qiangtang et l'Asie au Trias (orogénèse cimmérienne), et la collision entre l'Inde et l'Asie au début du Paléogène (orogénèse himalayenne). L'accrétion du bloc de Lhasa au Crétacé inférieur, bien que de plus faible ampleur, a également eu des implications non-négligeables qui seront discutées dans cette thèse. De nombreux autres blocs (Indochine, Simao, Sukhothai, Malaisie est, Sibumasu, Sumatra ouest, Burma ouest, etc.) ont également été accrétés le long de la marge de l'Asie du Sud-Est depuis le Permien et durant le Mésozoïque (Fig. 1.1.4 ; e.g. Carter & Clift, 2008 ; Sone & Metcalfe, 2008 ; Morley et al., 2013). Cette zone très complexe est largement en dehors du cadre de cette thèse et ne sera pas détaillée. Pour une synthèse, le lecteur peut consulter entre autres les travaux de Metcalfe (2006, 2013) (cf. Fig. 1.1.5).

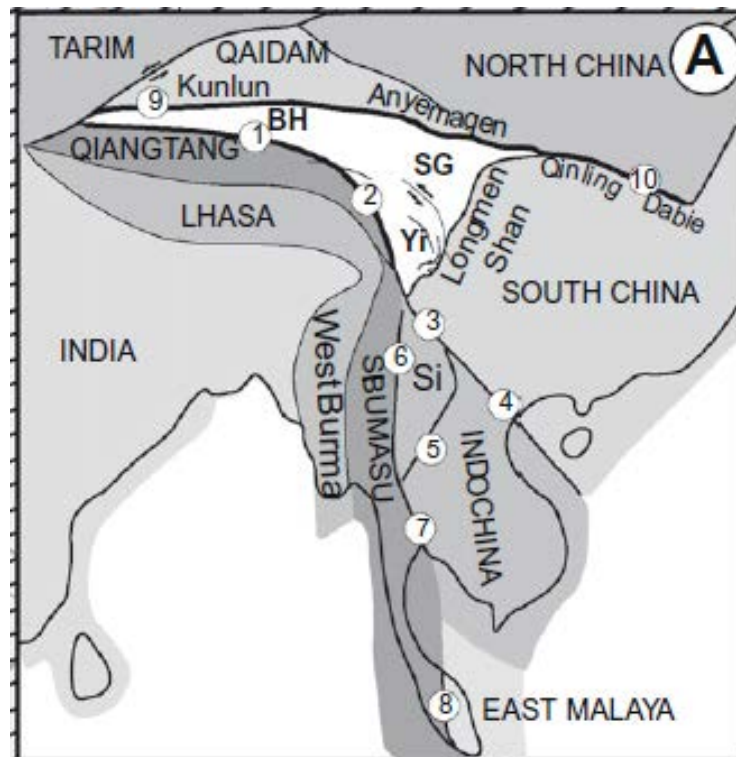


Figure 1.1.4 : Localisation des blocs constituant l'Asie actuelle.

BH = Bayan Har ; SG = Songpan-Garzê ; Y = Yidun ; Si = Simao (Roger et al., 2010).

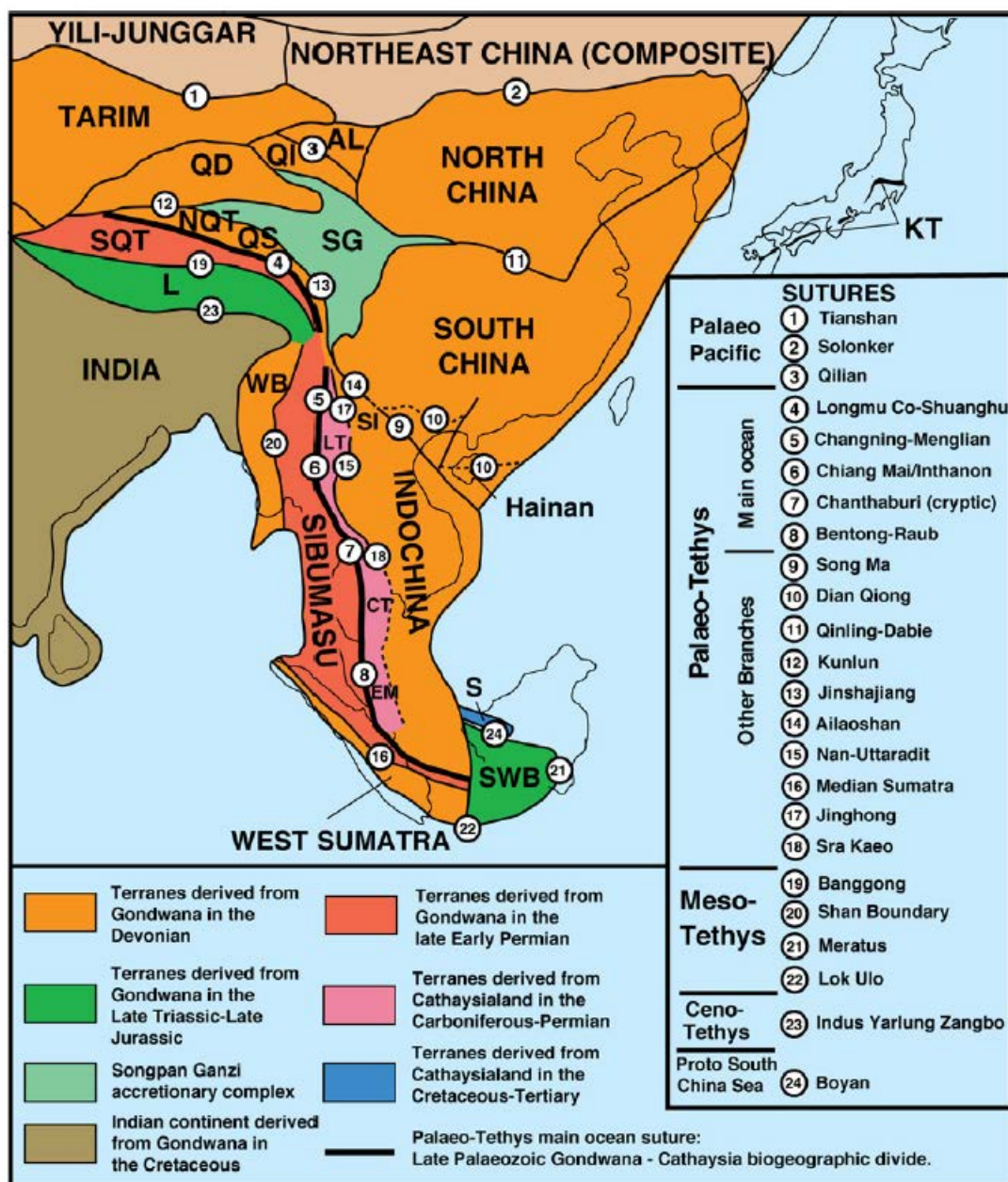


Figure 1.1.5 : Répartition des principaux blocs et sutures en Asie du Sud-Est. WB = Burma Ouest, SWB = Borneo Sud-Ouest, S = Semitau, L = Lhassa, SQT = Qiangtang Sud, NQT = Qiangtang Nord, QS = Qamdo-Simao, SI = Simao, SG = prisme d'accrétion de Songpan Garzê, QD = Qaidam, QI = Qilian, AL = Ala Shan, KT = Kurosegawa Terrane, LT = Lincang arc Terrane, CT = Chanthaburi arc Terrane, EM = East Malaya (Metcalf, 2013, d'après Metcalfe 2011 b).

A l'ouest de la CAO, une zone de subduction vers le nord se forme au Jurassique – Crétacé, le long d'une nouvelle marge continentale active, au sud des plaques des Pontides, du Trans-Caucase et d'Iran. A l'arrière de la zone de subduction, l'extension arrière-arc permet la création du bassin du Grand Caucase (Proto Bassin Sud-Caspien) à partir du Jurassique moyen à supérieur et dont l'étendue maximale est atteinte au Crétacé supérieur (Fig. 1.1.6 ; e.g. Brunet et al., 2003 ; Golonka, 2004). A la fin du Crétacé, les plaques du Caucase inférieur, Sanandaj-Sirjan et Makran ont été

accrétées aux plaques Iranienne et Afghane dans la région du Caucase – Mer Caspienne (e.g. Golonka, 2004).

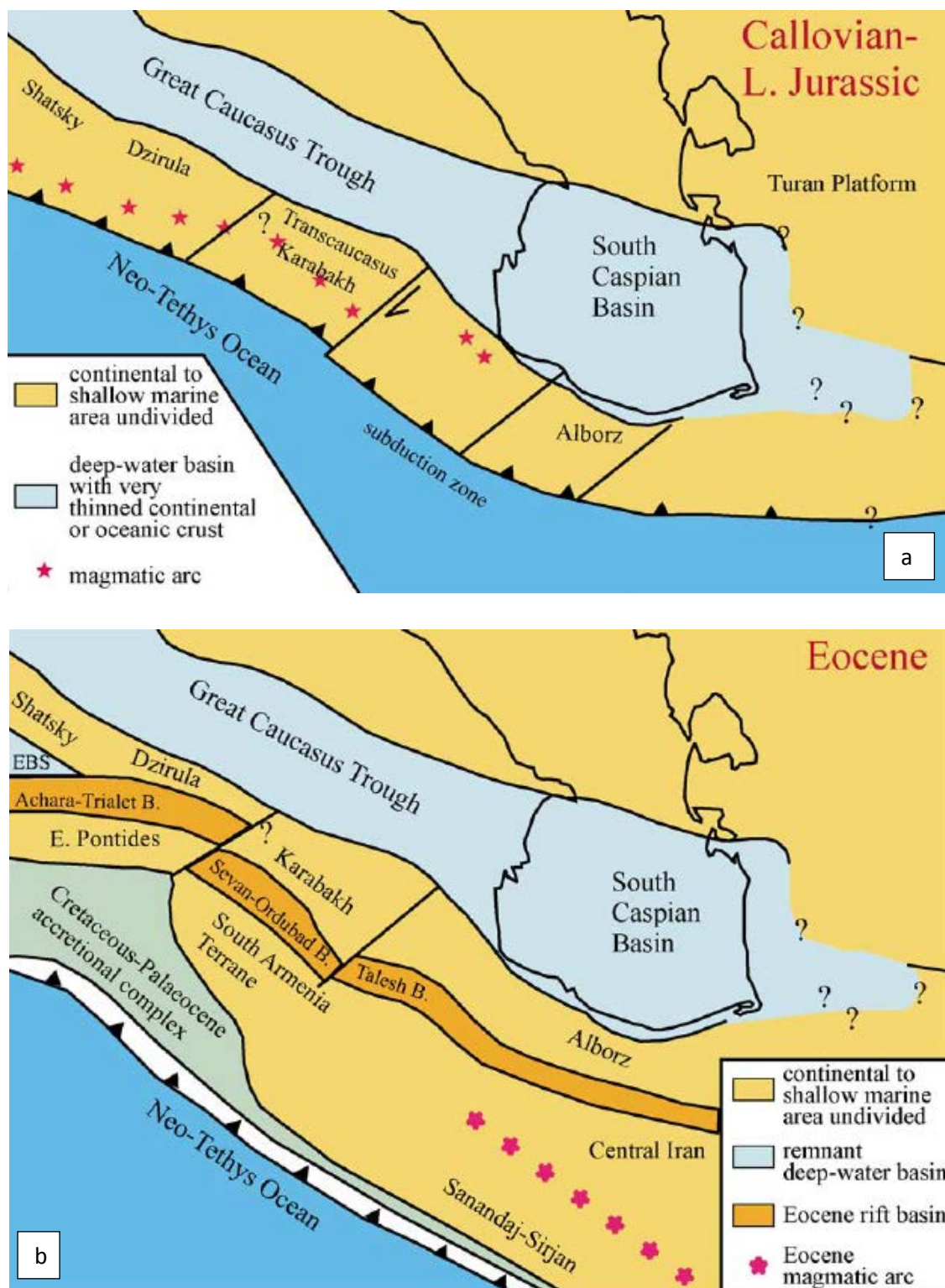


Figure 1.1.6 : Reconstruction paléogéographique de la zone sud Caspienne au Callovien – Jurassique supérieur (a) et à l'Eocène (b). EBS – Mer Noire est (Brunet et al., 2003).

Cette réorganisation des continents accompagnée de la formation de nouveaux reliefs, a certainement impliqué des changements de circulation des masses d'air pouvant entraîner des variations climatiques significatives (e.g. Hendrix et al., 1992 ; Dupont-Nivet et al., 2007). L'évolution des conditions climatiques sera détaillée dans les paragraphes qui suivent. Cependant, elles restent souvent approximatives pour la période du Mésozoïque, en raison de la difficulté à reconstruire un climat ancien, pour lequel les archives sont souvent peu ou mal conservées.

Deux événements majeurs marquent l'histoire tectonique Mésozoïque de l'Asie et méritent d'être expliqués ici afin de mieux comprendre leurs conséquences sur la structuration de la lithosphère asiatique. Il s'agit de la subduction et fermeture des océans Paléotéthys et Mongol-Okhotsk, respectivement au sud et au nord de la CAO.

1.1.1) Paléozoïque supérieur – Mésozoïque inférieur

1.1.1.1) Permien – Trias

1.1.1.1.2) La fermeture de l'Océan Mongol-Okhotsk

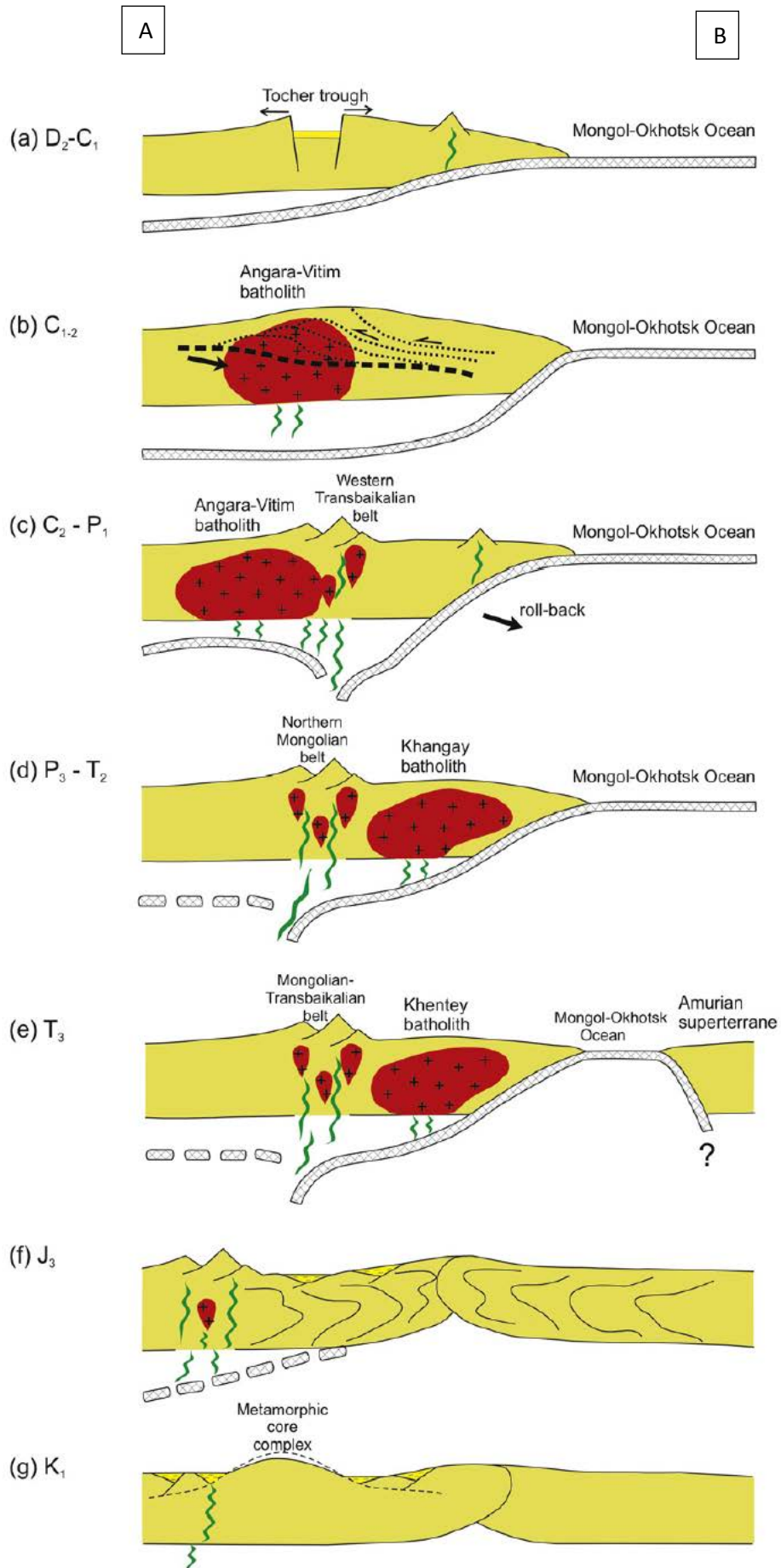
Au nord-est de la CAO, l'océan Mongol-Okhotsk est une branche de l'océan Paléo-Pacifique et se situe entre les continents Sibérie et Mongolie-Chine du Nord (e.g. Zonenshain et al., 1990 ; Zorin, 1999 ; Fig. 1.1.7). La lithosphère océanique de l'océan Mongol-Okhotsk fait l'objet d'une subduction continue sous le bloc Sibérie (vers le nord-ouest) probablement depuis le Dévonien, comme l'attestent les turbidites d'âge dévonien à carbonifère observées dans les bassins de Khangay et Khentey, ainsi que des reconstructions paléogéographiques du bloc Sibérie et de ceux alentours, basées sur la datation et l'étude géochimique des roches magmatiques (Fig. 1.1.8 ; e.g. Kravchinsky et al., 2002b ; Windley et al., 2007). La subduction du panneau plongeant est fortement ralentie en raison du faible angle de la subduction (Fig. 1.1.7). Ce ralentissement entraîne probablement une augmentation de la densité du panneau par déshydratation de la croûte, ce qui entraîne une augmentation de l'angle de subduction (Donskaya et al., 2013). Malheureusement aucune roche attestant de ces processus (roche de type éclogitique par exemple) n'a été observée. Au Carbonifère, l'augmentation de l'angle de subduction se traduit par un contexte tectonique devenu clairement compressif, entraînant l'épaississement de la croûte continentale Sibérienne, ainsi que la mise en place de granites (batholithe d'Angara-Vitim) typiques d'une zone de collision (Tsygankov et al., 2007). A la fin du Carbonifère et au début du Permien, le retrait du panneau plongeant en direction de l'océan entraîne l'extension de la lithosphère continentale et permet une contamination

mantellique de la croûte. Cette dernière présente de nombreuses intrusions magmatiques qui pourraient avoir une source mantellique (Jahn et al., 2009 ; Donskaya et al., 2013). La formation des granitoïdes alcalins et des roches volcaniques de la ceinture Transbaïkale ouest (Fig. 1.1.8) a donc lieu lors de l'extension de la croûte continentale. Entre la fin du Permien et le Trias moyen, la marge continentale active de Sibérie (au nord de la suture du Mongol-Okhotsk) est affectée par la formation du batholithe de Khangay, composé de granitoïdes calco-alcalins. Ils sont datés entre 261 ± 3 Ma et 241 ± 2 Ma par la méthode U-Pb sur zircons (voir nombreuses références dans Donskaya et al., 2013). La subduction se poursuit avec un angle plus plat, dit « normal » comparé à celui de la subduction carbonifère (Fig. 1.1.7).

A la fin du Trias, le batholithe de Khentey se forme au nord de la zone de suture du Mongol-Okhotsk, alors que des granitoïdes alcalins et des laves bimodales se forment en contexte extensif dans l'arrière-arc, au-dessus du panneau plongeant restant après la rupture de ce dernier. Van der Voo et al. (1999) ont montré grâce à la tomographie sismique, l'existence de restes du panneau plongeant sous le continent Sibérien, qu'ils supposent être d'âge jurassique.

Cette zone correspond à la chaîne Mongolo-Transbaïkale, située au nord des batholithes cités précédemment (Fig. 1.1.7). Cette organisation des complexes magmatiques montre l'existence d'une marge continentale active, se formant au-dessus d'une zone de subduction (Fig. 1.1.7, 1.1.8 ; Barbarin, 1999).

Figure 1.1.7 : Modèle schématique de l'évolution de la subduction de l'océan Mongol-Okhotsk. Localisation schématique (A-B) sur la carte (Fig. 1.1.8) (Donskaya et al., 2013).



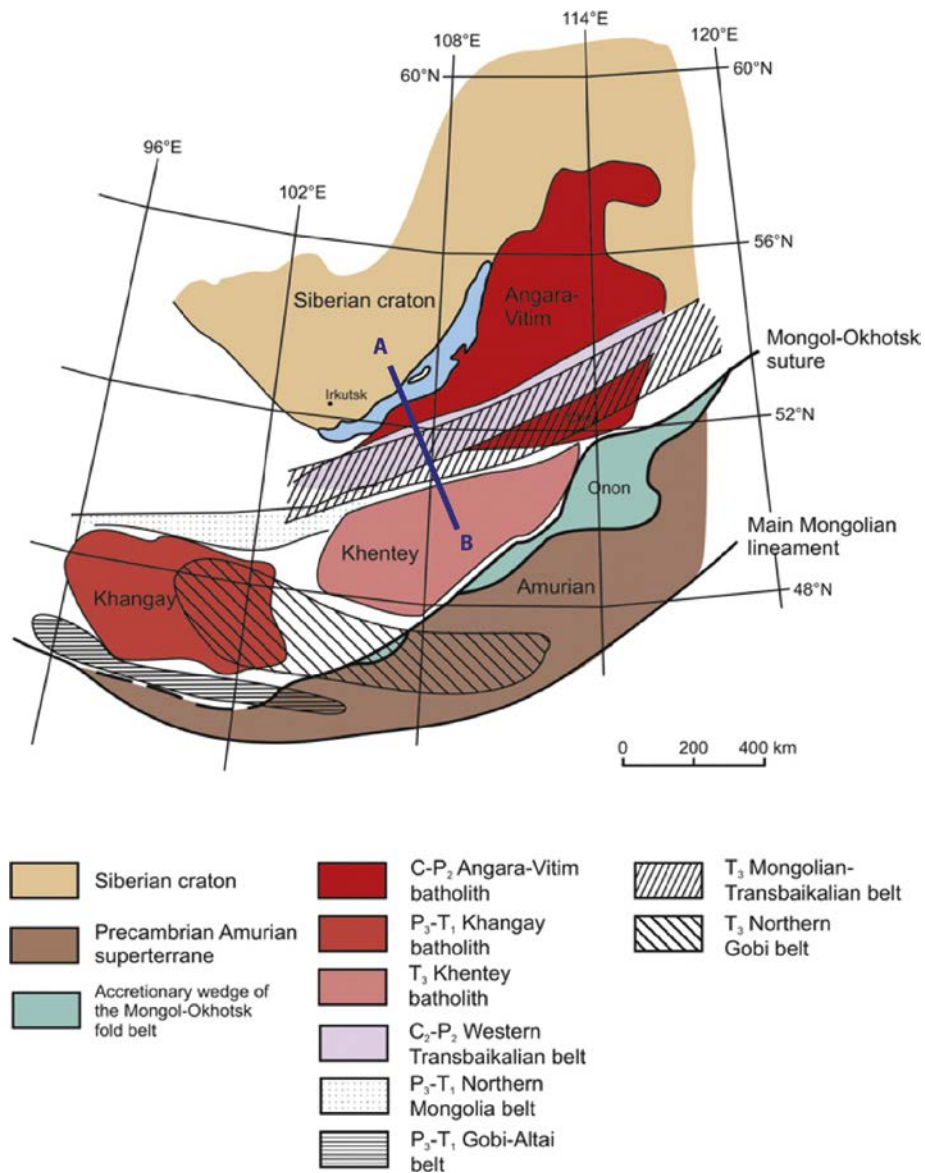


Figure 1.1.8 : Carte de la répartition des batholithes et ceintures volcano-plutoniques d'âge tardi-paléozoïques et mésozoïques dans la partie nord-est de la CAOB (Donskaya et al., 2013).

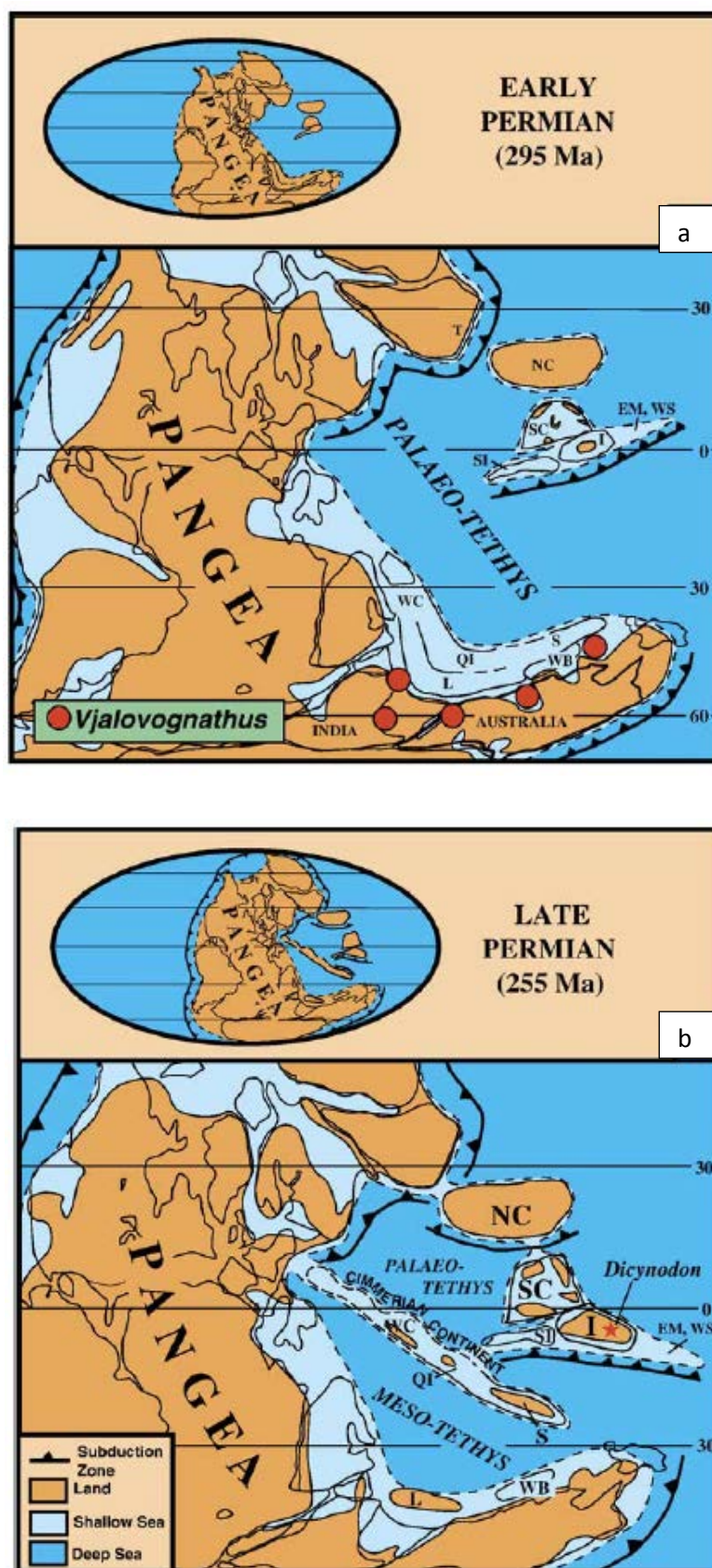


Figure 1.1.9 Reconstitution paléogéographique montrant la convergence des blocs cimmeriens (appelés ici « cimmerian continent ») et de la marge d'Asie du Sud-Est au Permien inférieur (a) et au Permien supérieur (b). SC= Chine du Sud; T= Tarim; I= Indochine; Em= Est Malaya; WS= Ouest Sumatra; NC= Chine du Nord; SI= Simao; S= Sibumasu; WB= Ouest Burma; QI= Qiangtang; L= Lhasa; WC= Continents ouest-cimmeriens (Metcalf, 2006).

1.1.1.1.2) Les épanchements basaltiques en Asie

Durant le Permien et au début du Trias, la mise en place de gigantesques épanchements basaltiques (LIP, *Large Igneous Province*) a lieu dans le bassin du Tarim (TLIP, 292 - 272 Ma, e.g. Zhang et al., 2008 ; Li et al., 2011 ; Yu et al., 2011 ; Zhang et al., 2012), puis dans les Emeishan en Chine du Sud (ELIP, 262 - 257 Ma, e.g. He et al., 2007 ; Guo et al., 2004 ; Shellnutt et al., 2012) et pour finir en Sibérie (SLIP, 252 – 248 Ma, e.g. Campbell et al., 1992 ; Renne et al., 1995 ; Kamo et al., 1996 ; Ivanov, 2007 ; Pirajno et al., 2010 ; Sharma, 2013) (Fig. 1.1.10). Ces mises en place de laves basaltiques se succèdent dans le temps et sont dues à l'activité de trois plumes mantelliques distincts (e.g. Pirajno et al., 2009 ; 2010). L'activité intense de ces trois plumes sur une période de 30-40 Ma pourrait indiquer la présence d'un « superplume » dans le manteau profond, qui aurait engendré une série de plumes secondaires (Courtilot et al., 2003).

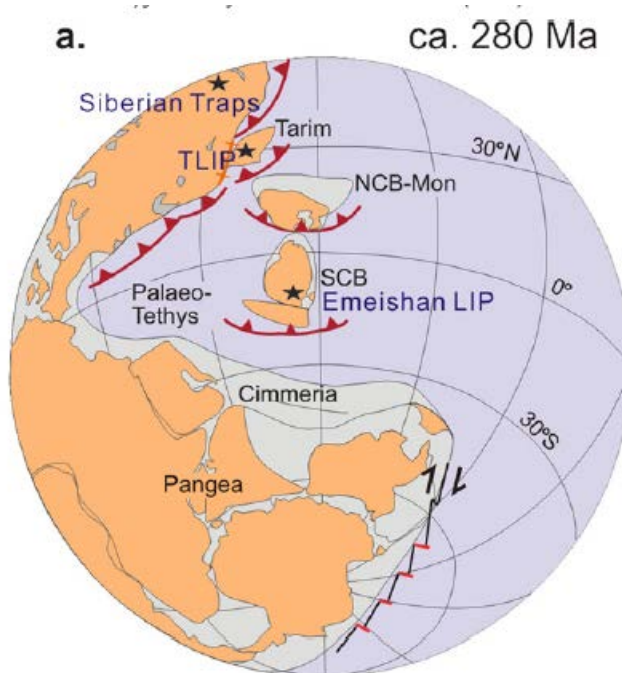


Figure 1.1.10 : Localisation des trois LIP permiennes en Asie : Tarim (TLIP), Emeishan (ELIP) et Sibérie (SLIP ou Siberian Traps). NCB-Mon = Bloc Chine du Nord – Mongolie ; SCB = Bloc Chine du Sud (Li Z. & al., 2011, modifié d'après Li & Powell, 2001).

1.1.1.1.2.1) Tarim

Les épanchements de laves basaltiques ayant eu lieu au Permien inférieur dans la partie centrale et ouest du bassin du Tarim (TLIP ou LIP de Bachu) couvrent plus de 250 000 km² (Fig. 1.1.11, e.g. Zhang et al., 2008). Le LIP du Tarim ne se limite pas au bassin du même nom. Des roches mafiques lui ont été associées dans l'Altay et le sud de la Sibérie, ainsi que dans le nord du bloc Chine du Nord (Fig. 1.1.12 ; e.g. Borisenko et al., 2006 ; Zhang et al., 2008 ; Su et al., 2012 et références associées ; Zhang & Zou, 2013).

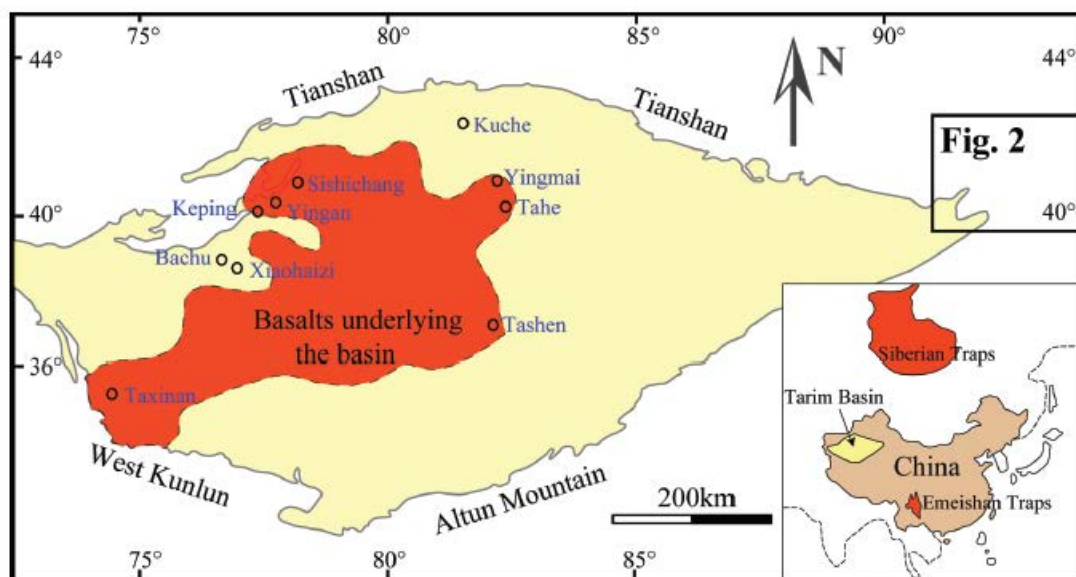


Figure 1.1.11 : Etendue minimale des épanchements basaltiques permien inférieur dans le Bassin du Tarim (Qin et al., 2011).

Dans le bassin du Tarim, les méthodes U–Pb sur zircon et ⁴⁰Ar/³⁹Ar sur roches totales ont été utilisées pour dater l'épanchement entre 292 et 272 Ma, avec 90% des éruptions ayant eu lieu entre 292 et 287 Ma (e.g. Li et al., 2011 ; Zhang et al., 2012). D'après les données de géochimie, il s'agit de basaltes alcalins dérivés d'une source mantellique de type basaltes d'île océanique (OIB - *ocean island basalt*), mis en place dans un régime tectonique extensif (e.g. Li et al., 2011).

On associe cet épanchement à la formation du rift du Beishan (au sud du bassin de Turfan, Fig. i), qui s'est créé dans le contexte de la subduction de l'océan Tian Shan nord. Ces deux événements ont pris place respectivement au début, et à la fin du Carbonifère (Su et al., 2011b ; Qin et al., 2011).

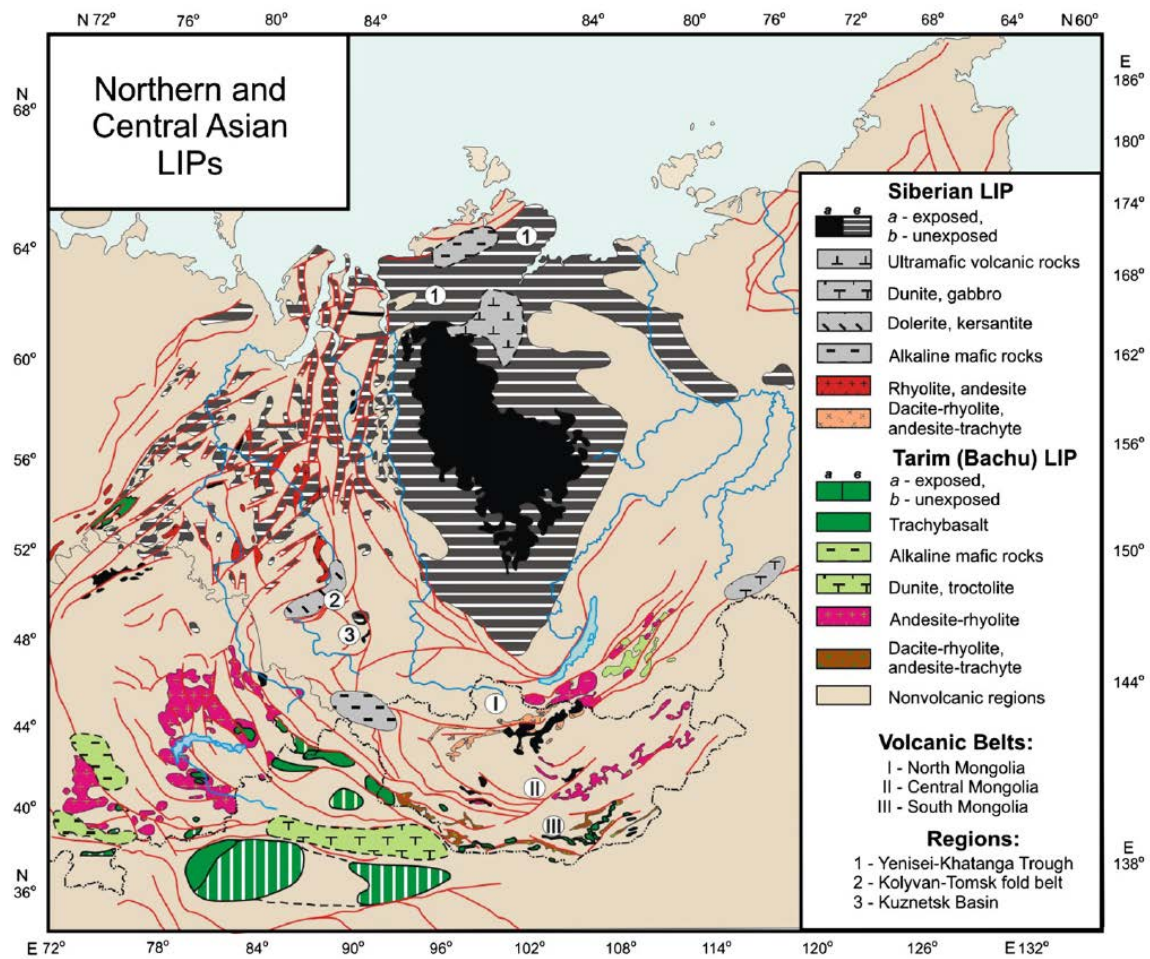


Figure 1.1.12 : Distribution des LIP du Tarim et de Sibérie (Pirajno et al., 2009).

1.1.1.1.2.2) Emeishan

L'épanchement basaltique des Emeishan (ELIP) couvre environ 300 000 km² dans le sud-ouest de la Chine, dans la partie est du plateau Tibétain et la partie ouest du bloc Chine du Sud (Fig. 1.1.13) (e.g. Chung & Jahn, 1995 ; Xu et al., 2001 ; Roger et al., 2010 ; Shellnutt, 2013).

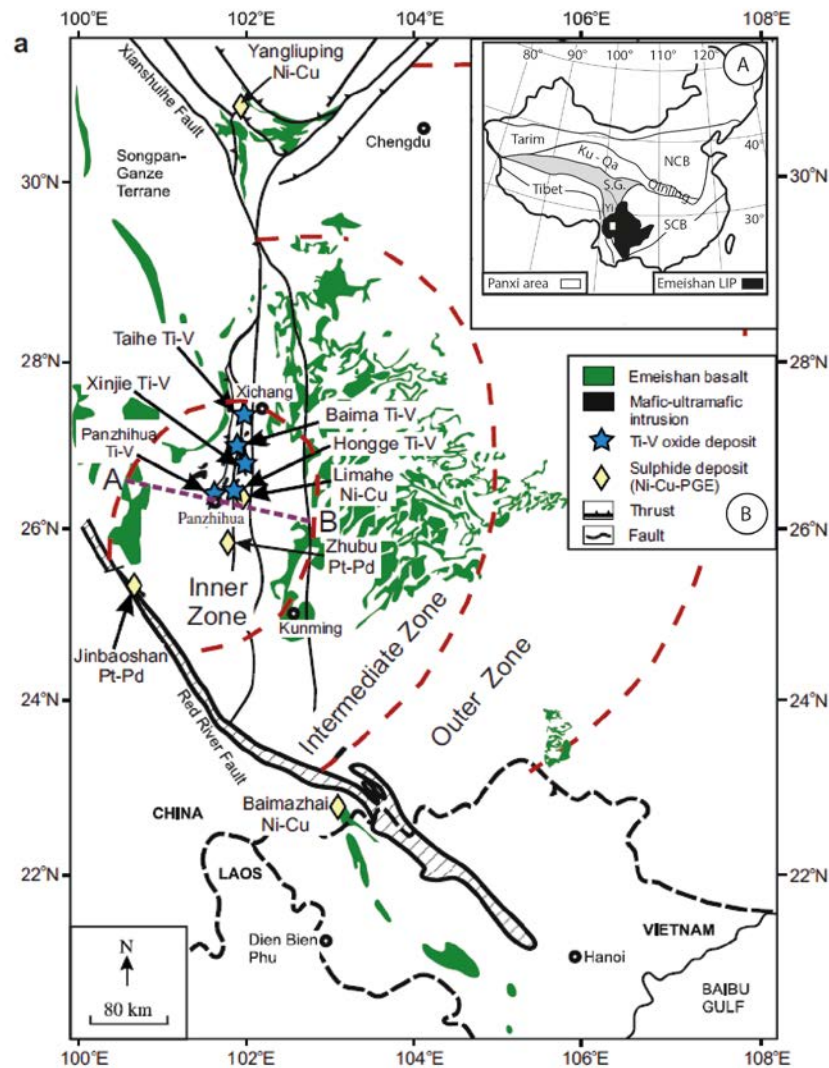


Figure 1.1.13 : Localisation des épanchements basaltiques des Emeishan (*Emeishan LIP*), à l'échelle de l'Asie (A) représentés en noir. Ku-Qa = Kunlun – Qaidam ; NCB = Bloc Chine du Nord ; SCB = Bloc Chine du Sud ; SG = Songpan-Garze ; Yi=Yidun (Roger et al., 2010) – B. à l'échelle régionale (Shellnutt, 2013).

Les derniers âges obtenus par datations U/Pb sur zircons indiquent une mise en place des basaltes entre 250 et 260 Ma, précédés par quelques dykes vers 262 ± 3 Ma (e.g. Zhou et al., 2002 ; Guo et al., 2004 ; Zhong et al., 2007 ; Shellnutt and Zhou, 2008), plus précisément entre 257.6 ± 0.5 Ma et 259.6 ± 0.5 Ma d'après les derniers résultats obtenus avec la ID-TIMS sur zircon (Shellnutt et al., 2012 ; Shellnutt, 2013). Les données géologiques, paléomagnétiques et géochronologiques s'accordent à dire que la durée d'éruption ne dépassa guère 3 Ma (Shellnutt, 2013). Deux modèles tentent d'expliquer ces épanchements basaltiques, soit comme la conséquence d'un mécanisme intra-lithosphérique (ouverture du Rift du Panxi ; e.g. Ali et al., 2005 et références associées), soit par la perforation de la lithosphère du bloc Chine du Sud par un plume mantellique (e.g. Xu et al., 2001, 2004; Xiao et al., 2004a; Song et al., 2004). D'autre part, l'éruption vers 260 Ma est synchrone de la

formation du Rift du Panxi le long de la marge ouest du bloc Chine du Sud, à l'arrière de la zone de subduction de Yushu-Batang. L'extension de la lithosphère entraîne la formation d'un point triple. Le micro-bloc de Yidun se détache du bloc principal (Chine du Sud) pour ouvrir le bassin océanique de Garzê-Litang (qui se prolonge au sud par le Rift du Panxi), qui constitue une branche de la Paléotéthys (Fig. 1.1.14 ; e.g. Mattauer et al., 1992 ; Reid et al., 2005 ; Roger et al., 2010).

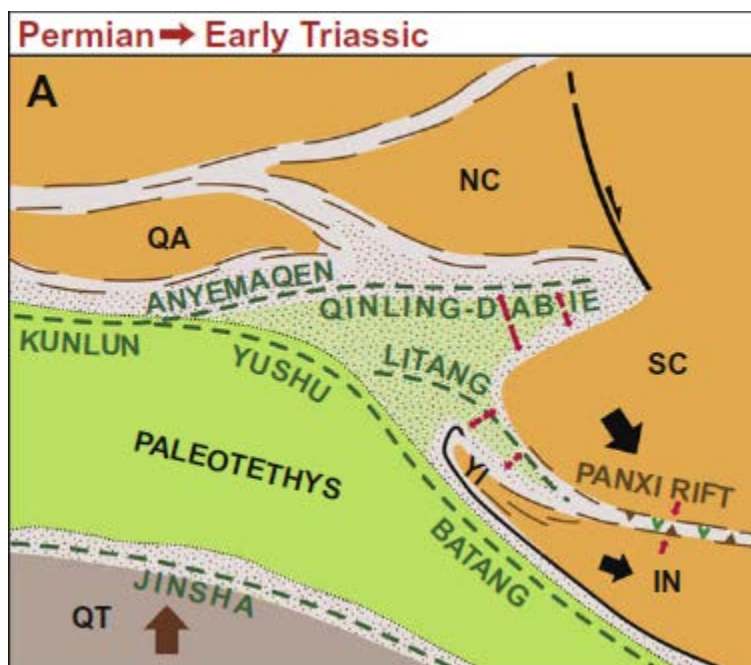


Figure 1.1.14 : Carte paléogéographique du Tibet Nord à la limite Permien – Trias, lors de la formation du Rift du Panxi.
Les futures zones de subduction sont représentées par les tirets verts. QA = bloc de Qaidam ; NC = bloc Chine du Nord ; SC = bloc Chine du Sud ; IN = bloc Indochine ; QT = bloc de Qiangtang ; YI = bloc de Yidun (ou Litang-Batang) (Roger et al., 2008, 2010).

1.1.1.1.2.3) Sibérie

L'épanchement basaltique (SLIP) qui a lieu en Sibérie à la limite Permien-Trias est le plus grand avec une étendue de $4 * 10^6$ km². Il est également appelé « Trapps de Sibérie » et couvre une grande partie du craton sibérien et du bassin ouest-sibérien. Il atteint la chaîne de l'Oural à l'ouest et le Kazakhstan au sud (Fig. 1.1.15 ; Ivanov, 2007). Sa mise en place a eu lieu entre 252 et 248 Ma, avec un pic d'activité magmatique d'un Ma vers 251 Ma, soit à la limite Permien-Trias (e.g. Campbell et al., 1992 ; Renne et al., 1995 ; Kamo et al., 1996 ; Sharma, 2013 ; Ivanov, 2007 ; Pirajno et al., 2009). Plus récemment, il a été montré que deux pulses magmatiques ont vraisemblablement eu lieu (Ivanov et al., 2013). Ces épanchements basaltiques sont dus à l'activité d'un plume mantellique (e.g. Richards et al., 1989). L'éruption du LIP a lieu dans un contexte extensif et est étroitement liée à un système de rifts dans le bassin Ouest-Sibérien (e.g. Courtillot et al., 1999 ; Allen et al., 2006 ; Pirajno et al., 2009).

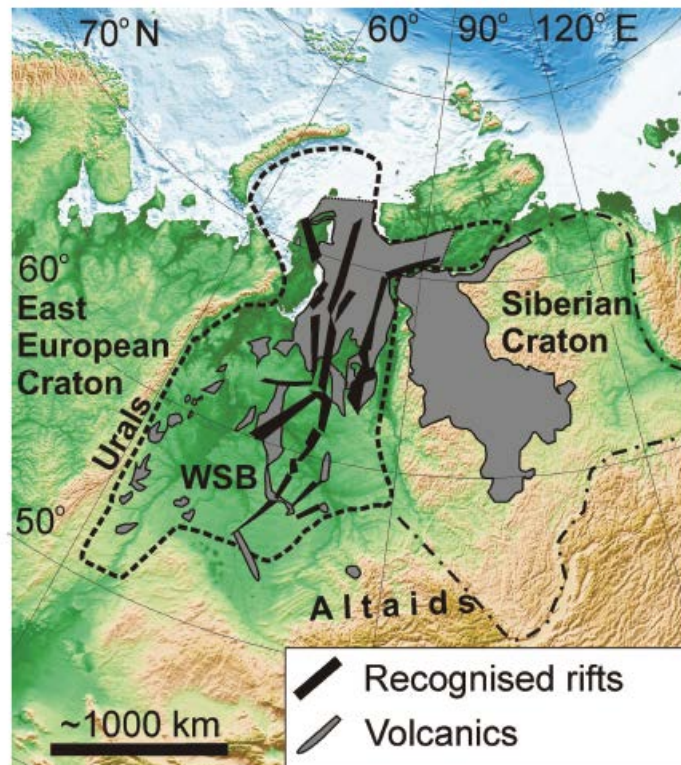


Figure 1.15 : Carte de situation du LIP de Sibérie et du système extensif associé. WSB = Bassin Ouest-Sibérien (Allen et al., 2006).

L'activité magmatique liée au plume entraîne l'amincissement de la lithosphère et accentue le rifting, ce qui a pour conséquence d'augmenter la subsidence dans le Bassin Ouest-Sibérien (Holt et al., 2012). Les dépôts sédimentaires contemporains sont ainsi préservés dans le sud-est du Bassin Ouest-Sibérien (e.g. Davies et al., 2010) et dans les rifts qui se forment à la même époque (Buslov et al., 2010).

1.1.1.1.3) L'Asie Centrale

Cette période correspond également à des mouvements de translation du bloc Yili-Junggar en direction de l'est par rapport au Tarim (1160 +/-380 km) le long de failles décrochantes dextres situées au nord, au sud et dans la chaîne du Tian Shan. Ces failles sont des structures majeures dans l'histoire du Tian Shan et sont sans doute à l'origine d'une topographie significative. L'activité transtensive engendre la formation de bassins en pull-apart permien (e.g. les bassins de Yili, Turfan ; e.g. Allen et al., 1995 ; Wang et al., 2009 ; Charvet et al., 2011), dont l'extension cesse à la fin du Permien – début du Trias et qui constitueront des centres de dépôt pour les sédiments mésozoïques (e.g. Laurent-Charvet et al., 2002; Shu et al., 2002; Wang et al., 2007a, de Jong et al., 2009). La

topographie permienne est probablement réactivée par les effets de l'orogénèse cimmérienne, telle que le suggère une phase de refroidissement enregistrée par les données de thermochronologie basse-température au Jurassique inférieur (cf. Dumitru et al., 2001 ; Jolivet et al., 2010, 2013c).

1.1.1.2) Trias

Au sud du continent, les blocs cimmériens commencent à s'éloigner de la marge himalayo-australienne du Gondwana au début du Permien, ouvrant ainsi l'océan Mésotéthys et initiant la fermeture de l'océan Paléotéthys (Fig. 1.1.9 ; Metcalfe 2006, Metcalfe, 2006, 2013).

Au Trias, l'histoire tectonique de l'Asie est fortement marquée par l'orogénèse cimmérienne. Celle-ci implique différents blocs et a lieu depuis la fin du Permien jusqu'au début du Jurassique. Dans la partie sud-est de l'Asie, il s'agit principalement des interactions au Trias entre le bloc de Sibumasu (prolongement est du bloc de Qiangtang) et le bloc Indochine (comprenant les blocs de Simao, Sukhothai et de la Malaisie est, Fig. 1.1.5) (Fig. 1.1.16 ; e.g. Carter & Clift, 2008 ; Sone & Metcalfe, 2008 ; Roger et al., 2008, 2010 ; Metcalfe, 2006, 2013 ; Morley et al., 2013). C'est ce dernier événement qui est couramment appelée « orogénèse indosinienne » (Deprat, 1914 ; Fromaget 1941). Une confusion règne autour de ce terme utilisé depuis plusieurs décennies plus au nord au Tibet. Récemment, il a été proposé de réserver l'utilisation de ce terme à l'activité anté-norienne définie au Vietnam (Roger et al., 2008, 2010). Dans la partie nord-est du Tibet actuel, une forte activité tectonique liée à la fermeture de la Paléotéthys affecte les blocs du Qaidam, Qiangtang, Kunlun, Chine du Nord et Chine du Sud du Trias au Jurassique supérieur (Fig. 1.1.16, e.g. Zou et al., 1984 ; Sengör, 1985; Mattauer et al., 1992 ; Jolivet et al., 2001 ; Roger et al., 2004, 2008, 2011 ; Reid et al., 2005 ; Zhang et al., 2006a ; Pullen et al., 2008).

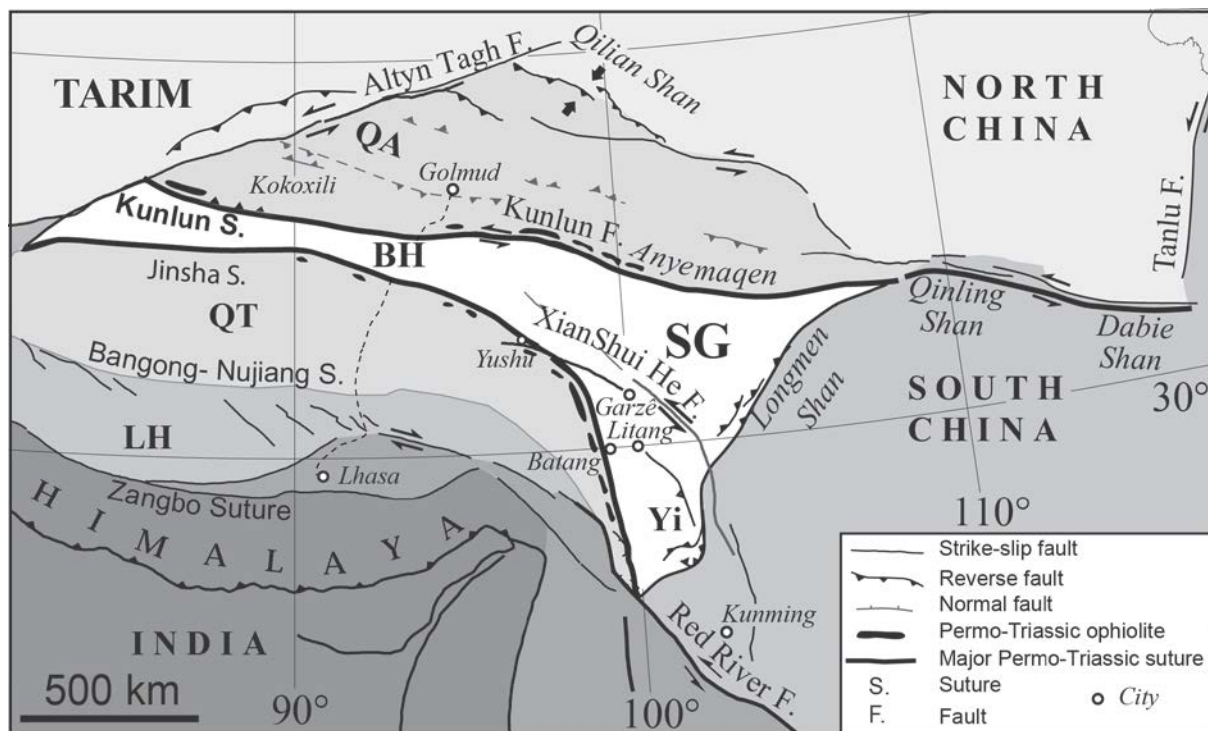


Figure 1.1.16 : Carte représentant les blocs et principales sutures impliquées dans l'orogénèse triasique. SG = chaîne de Songpan-Garzê, BH = terrane de Bayan Har, Yi = bloc de Yidun (ou Litang-Batang), LH = bloc de Lhasa, QT = bloc de Qiangtang, QA = bloc de Kunlun-Qaidam (Roger et al., 2011).

Les blocs Indochine et Chine du Sud ont sans doute été accrétés à la limite Permien – Trias, le long de la suture de Song Ma (Roger et al., 2014). Concernant la collision entre les blocs Chine du Nord et Chine du Sud, deux écoles s'affrontent et proposent soit un âge permien (e.g. Faure et al., 2009 ; Li & Li, 2007), soit triasique (e.g. Dong et al., 2012) cf. explication dans Metcalfe 2013. La conséquence de cette collision est l'orogénèse des Qinling-Dabie (qui par ailleurs constitue une des sources de sédiments du prisme de Songpan-Garzê) (Roger et al., 2010).

L'âge de la collision entre les blocs Chine du Nord et Chine du Sud est très largement débattu, avec des interprétations allant du Paléozoïque moyen jusqu'au Mésozoïque inférieur à supérieur (e.g. Mattauer et al., 1985; Faure et al., 2008; Roger et al., 2010). En réalité, il semblerait que deux épisodes de collision aient eu lieu, le premier au Paléozoïque moyen et le second au Mésozoïque inférieur, séparés par un court épisode de rifting décrit ci-après (e.g. Enkelmann et al., 2006).

De manière générale, la subduction de la croûte océanique (au nord du bloc Chine du Sud) sous le bloc Chine du Nord entraîne la fermeture progressive de l'océan Paléotéthys. Une partie de cette histoire est enregistrée dans l'évolution de l'océan de Songpan Garzê, qui formait le domaine est de l'océan Paléotéthys. Au Permien, Songpan Garzê était divisé en deux sous-bassins : Songpan-Garzê et Garzê-Litang (Fig. 1.1.17 ; Roger et al., 2008). Le bassin de Songpan-Garzê s'est développé vers l'est en raison de mouvements entre les blocs Chine du Nord et Chine du Sud (Zhang et al., 2006b). Le bassin de Garzê-Litang s'est formé suite au rifting sur la marge du bloc Chine du Sud, qui a permis l'ouverture permienne du rift de Panxi au sud-est (e.g. Zhou et al., 2002 ; Roger et al., 2008 ;

Shellnutt & Zhou, 2008). Cette ouverture pourrait être liée entre autres à la présence du point chaud dans les Emeishan, ou à la rotation du bloc Indochine (e.g. Roger et al., 2010). De plus, des séries basaltiques permienes apparentées aux épanchements dans les Emeishan (cf. 1.1.1.1.2.2) sont présentes dans la zone de Songpan-Gârze (e.g. Roger et al., 2010 et références associées). Trois zones de subduction étaient actives durant le Trias pour fermer l’océan Paléotéthys : Kunlun-Anyemaqen au Nord, Jinsha au Sud et Yushu-Batang à l’est de l’océan (Fig. 1.1.16, 1.1.17 et 1.1.18 ; e.g. Yang et al., 1996 ; Konstantinovskaia et al., 2003 ; Roger et al., 2008 ; 2010 ; 2011). Durant le Trias, l’activité est synchrone le long des trois zones de subduction et nourrit un gigantesque prisme d’accrétion, qui subsiste aujourd’hui dans la zone triangulaire (en blanc sur la Fig. 1.1.16) située entre les blocs du Kunlun-Qaidam au Nord, Qiangtang au Sud-Ouest et Chine du Sud à l’Est (Fig. 1.1.17). Elle comprend d’Ouest en Est les blocs de Bayan Har (BH) et Songpan-Garzê (SG). Ce prisme où l’épaisseur des sédiments atteint jusqu’à 25 km a été largement déformé, et la croûte épaissie atteint localement plus de 50 km d’épaisseur (e.g. Malavieille et al., 1991 ; Calassou, 1994 ; Jolivet et al., 2003 ; Vergne et al., 2003 GRL ; Harrowfield & Wilson, 2005 ; Wang et al., 2007c ; Zhang et al., 2011a). Le prisme chevauche la marge du bloc Chine du Sud au Trias supérieur (e.g. Mattauer et al., 1992 ; Harrowfield & Wilson, 2005). Les zones de subduction de Kunlun-Anyemaqen et Jinsha fonctionnent avec des directions de vergence opposées, ce qui entraîne la fermeture du domaine océanique (Fig. 1.1.18). Le principal moteur de cette fermeture est le déplacement vers le nord du bloc de Qiangtang. Toutefois, la fermeture de l’océan Paléotéthys semble être incomplète à l’est. En effet, Roger et al. (2010, 2011) supposent que la forme triangulaire de la zone de Songpan-Garzê, associée au prisme d’accrétion très épais, a empêché la collision continent-continent. Par conséquent, le socle de la partie nord-est de la zone, sous le prisme d’accrétion, pourrait être composé de croûte océanique (Roger et al., 2010). L’épaississement du prisme d’accrétion est à l’origine de la mise en place de granites syn- à tardi-orogéniques (220 -200 Ma) (Roger et al., 2004 ; Zhang et al., 2006b), ainsi que de granites post-orogéniques (200-150 Ma) (Roger et al., 2004). De manière générale, ce prisme fait l’objet d’intrusions granitiques d’origines et d’âges variés (cf. discussion et références dans Roger et al., 2008 ; 2010).

Au Trias supérieur – Jurassique inférieur, le batholithe du Kunlun (220-200 Ma) se met en place dans le contexte de la subduction de Kunlun-Anyemaqen. La principale phase de formation des prismes orogéniques du Kunlun et de Qinling a lieu au Trias supérieur (e.g. Harris et al., 1988 ; Roger et al., 2003, 2008). Au sud de l’océan Songpan-Garzê, la déformation le long de la zone de subduction de Jinsha est marquée par le métamorphisme de type schistes bleus (220 Ma), qui indique une subduction continue (Kapp et al., 2000 ; Pullen et al., 2008).

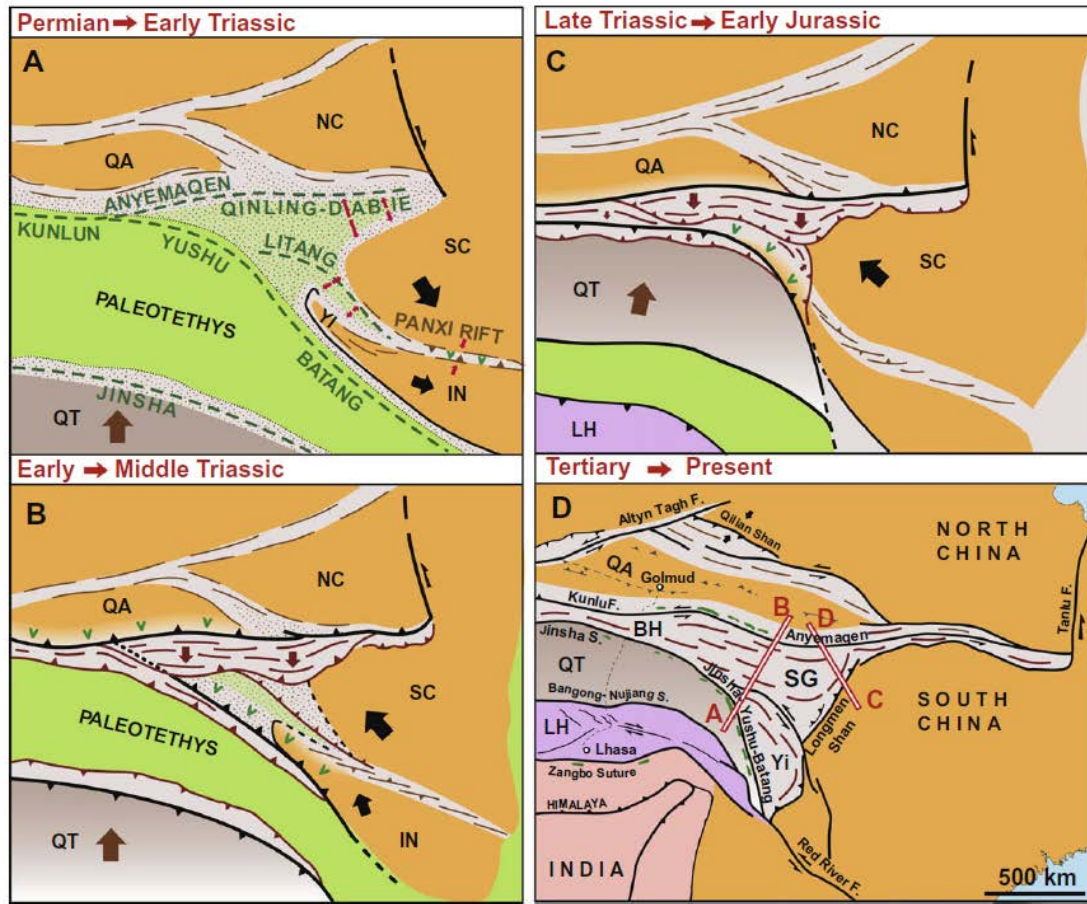


Figure 1.1.17 : Cartes schématiques synthétisant l'évolution tectonique du Nord Tibet, depuis le Permien jusqu'au Tertiaire. Yi = bloc de Yidun (ou Litang-Batang), IN = bloc Indochine, QT = bloc de Qiangtang, QA = bloc de Qaidam, NC = bloc Chine du Nord, SC = bloc Chine du Sud. AB et CD représentent la position des coupes évolutives de la Figure 1.1.18 (Roger et al., 2010).

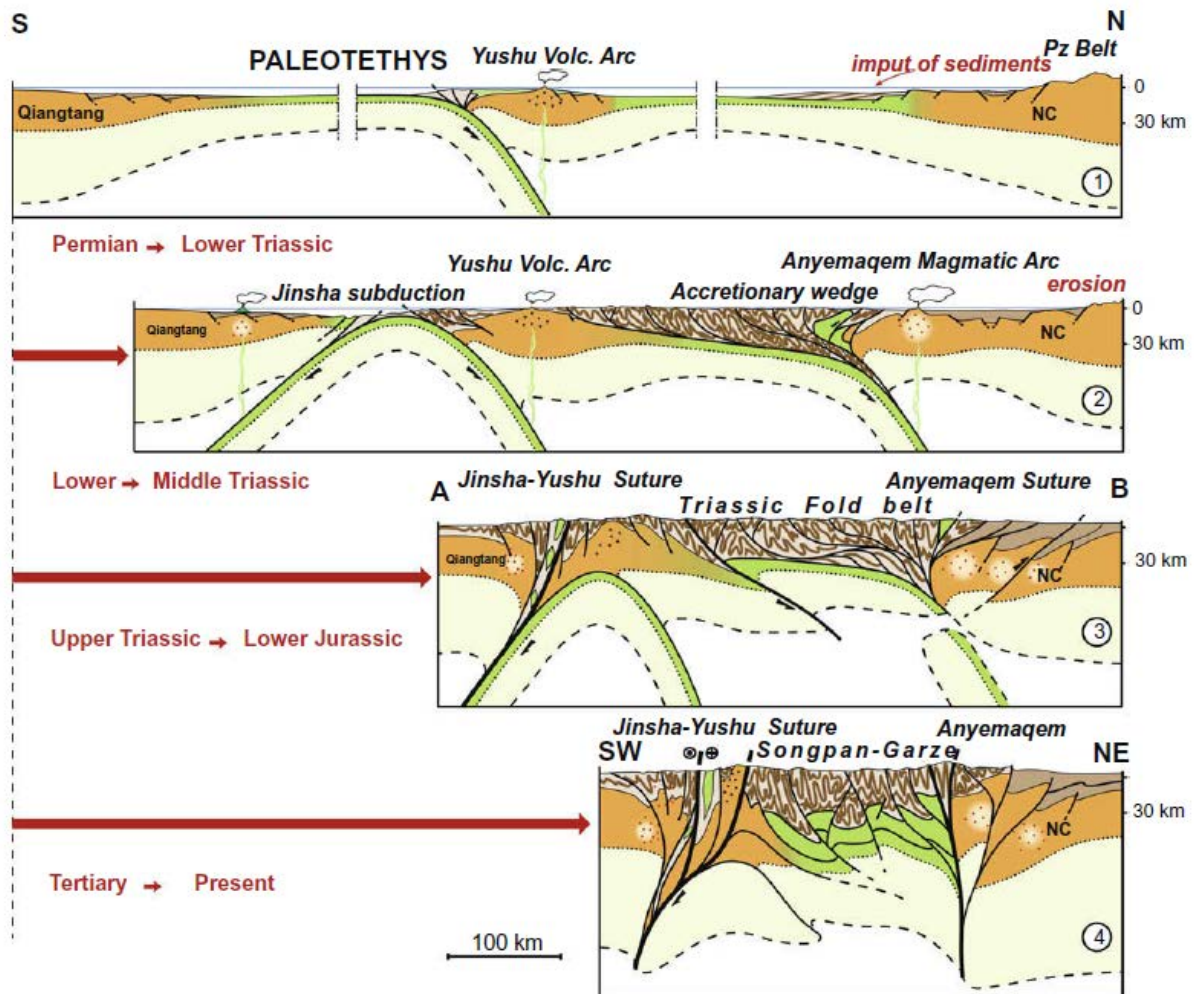


Figure 1.1.18 : Modèle de formation et d'évolution de la chaîne triasique du Permien au Tertiaire. (1) Début de la subduction de l'océan Paléotéthys au Permien. (2) Fermeture de l'océan Paléotéthys et subduction à double vergence, accompagné de la formation du prisme d'accrétion et de la mise en place du batholithe du Kunlun (220-200 Ma). (3) Fin de la subduction et déformation du prisme. (4) Fort raccourcissement dus à la collision Inde-Asie et subduction intracontinentale le long de la zone réactivée de Jinsha (Roger et al., 2008, 2010).

A la fin du Trias, la convergence des continents mène à la fermeture définitive (du moins dans la partie ouest) de l'océan Paléotéthys et à l'accrétion du bloc de Qiangtang le long de la suture de Jinsha (e.g. Mattauer et al., 1992 ; Kapp et al., 2000 ; Golonka, 2007 ; Pullen et al., 2008 ; Roger & al., 2010, 2011 ; Metcalfe, 2013) (Fig. 1.1.17). La fin de l'orogénèse est caractérisée par les dépôts continentaux tardi-triasiques légèrement déformés, qui reposent de manière discordante sur les séries marines du Trias moyen intensément déformées (Calassou, 1994). Dans la région de Songpan-Gârze, l'exhumation post-triasique est quasi-inexistante. La quasi-absence de dépôts post-triasiques sur la carte géologique de la zone (C.I.G.M.R., 1991), ainsi que leur présence dans les bassins alentours indiquerait un haut topographique dans cette région à la fin du Trias. Il est très difficile

d'estimer l'altitude de cette structure, cependant, elle ne peut être très haute, puisque le prisme n'est pas érodé (Roger et al., 2010).

Sur la bordure nord du Tibet actuel, les données de traces de fission sur apatite et les datations $^{39}\text{Ar}/^{40}\text{Ar}$ enregistrent l'exhumation due à la collision du bloc du Qiangtang, qui a lieu au Trias moyen à supérieur dans un système en compression (e.g. Jolivet et al., 1999 ; 2001 ; Sobel et al., 2001).

Dans la partie ouest de la Paléotéthys, la fermeture de l'océan et la collision des blocs Qamdo-Simao (QS, Fig. 1.1.5) et Chine du Sud engendre l'orogénèse de Jinshajiang (Zi et al., 2013). De façon plus générale, le lecteur intéressé peut se référer au travail de Angiolini et al. (2013).

Plus au sud, la déformation ductile qui affecte la région de l'actuel Vietnam et le métamorphisme associé sont bien antérieurs à la déformation du Tibet et sont datés à 250-240 Ma (datations U/Pb et Ar/Ar). De plus, la déformation a principalement été accommodée le long des grandes failles décrochantes (e.g. Lepvrier et al., 1997, 2004, 2008; Carter et al., 2001 ; Roger et al. 2014), mais aussi localement par la formation de nappes tectoniques, suggérant un raccourcissement significatif (Roger et al., 2012, 2014).

Au Trias, l'activité tectonique due à l'accrétion du bloc de Qiangtang permet la réactivation des principales structures tectoniques paléozoïques en Asie Centrale. Elle entraîne une phase d'exhumation majeure en Asie Centrale, qui s'accompagne de la création d'une topographie significative (e.g. Hendrix et al., 1992 ; Dumitru et al., 2001 ; Vassalo et al., 2007 ; Buslov et al., 2008 ; Jolivet et al., 2010). Cependant, il est difficile aujourd'hui de différencier la contribution de la tectonique permienne, de celle de la réactivation triasique dans la topographie mésozoïque (Jolivet et al., 2013c).

1.1.2) Mésozoïque Moyen - Supérieur

Dans la région du Mongol-Okhotsk, au début du Jurassique moyen, les derniers sédiments marins laissent place aux dépôts continentaux dans la région du Transbaikal (Fig. 1.1.19). Ce changement est accompagné d'une baisse de l'activité magmatique liée à la subduction de la lithosphère de l'océan Mongol-Okhotsk, et indique très probablement la fin de la subduction et le début de la collision entre la Sibérie et la Mongolie (Fig. 1.1.20 ; e.g. Zorin, 1999 ; Kravchinsky et al., 2002c ; Cogné et al., 2005 ; Jolivet et al., 2009, 2013a, 2013b ; Donskaya et al., 2013). Du fait de la

rotation horaire de la Mongolie par rapport à la Sibérie stable, la fermeture de l'océan Mongol-Okhotsk est diachrone d'ouest en est et entraîne une fermeture en ciseaux de cet océan (e.g. Cogné et al., 2005). La fermeture est définitive dans sa partie ouest au Jurassique inférieur à moyen (Fig. 1.1.20 ; Zorin, 1999) ou au Jurassique supérieur - Crétacé inférieur (Kravchinsky et al., 2002a ; Cogné et al., 2005). Des signes d'exhumation et de phases de refroidissement au Jurassique supérieur - Crétacé inférieur sont enregistrés par les données de thermochronologie et de sédimentologie dans les régions des Sayan, du Baikal et de la chaîne de Patom (e.g. Le Heron et al., 2008 ; Van der Beek et al., 1996 ; Jolivet et al., 2009 ; 2013b), ainsi que dans le nord-est de l'Altay (De Grave and Van den haute, 2002 ; De Grave et al., 2008). Cette phase d'exhumation semble être associée à la collision entre la Sibérie et la Mongolie et à la formation d'une chaîne de montagnes. L'« orogène » du Mongol-Okhotsk, tel que certains auteurs l'évoquent, se formeraient à ce moment-là (Fig. 1.1.11-f ; e.g. Donskaya et al., 2013). Pourtant sans trace de métamorphisme et sans évidence de racine crustale, il est difficile de parler d'une chaîne de montagnes (Jolivet et al., 2009). Dans la région du Transbaikal qui se trouve à proximité de la zone de suture, aucun métamorphisme relatif à une quelconque collision n'a pu être observé (Zorin, 1999). De la même manière, les résultats de thermochronologie basse-température (AFT) dans le sud-est de la chaîne des Sayan Est n'enregistrent aucune accélération du refroidissement à la fin du Jurassique – début du Crétacé (Jolivet et al., 2013b). Les sédiments crétacés dans le Bassin Ouest-Sibérien et dans la région Transbaïkale ne présentent ni dépôt grossier, ni discordance, ni variation majeure et ne semblent donc pas être le produit de l'érosion d'une chaîne de montagnes (e.g. Davies et al., 2010 ; Jolivet et al., 2009, 2013b). L'absence de critère indiquant une forte érosion et exhumation montre que cette « chaîne » devait être à relief limité et non associée à une déformation compressive intense. Récemment, Daoudene (2011) a proposé un modèle géodynamique pour expliquer une possible collision sans formation de chaîne de montagnes, impliquant une lithosphère anormalement chaude et une subduction à double vergence (cf. 1.1.3.1 Jurassique - Crétacé).

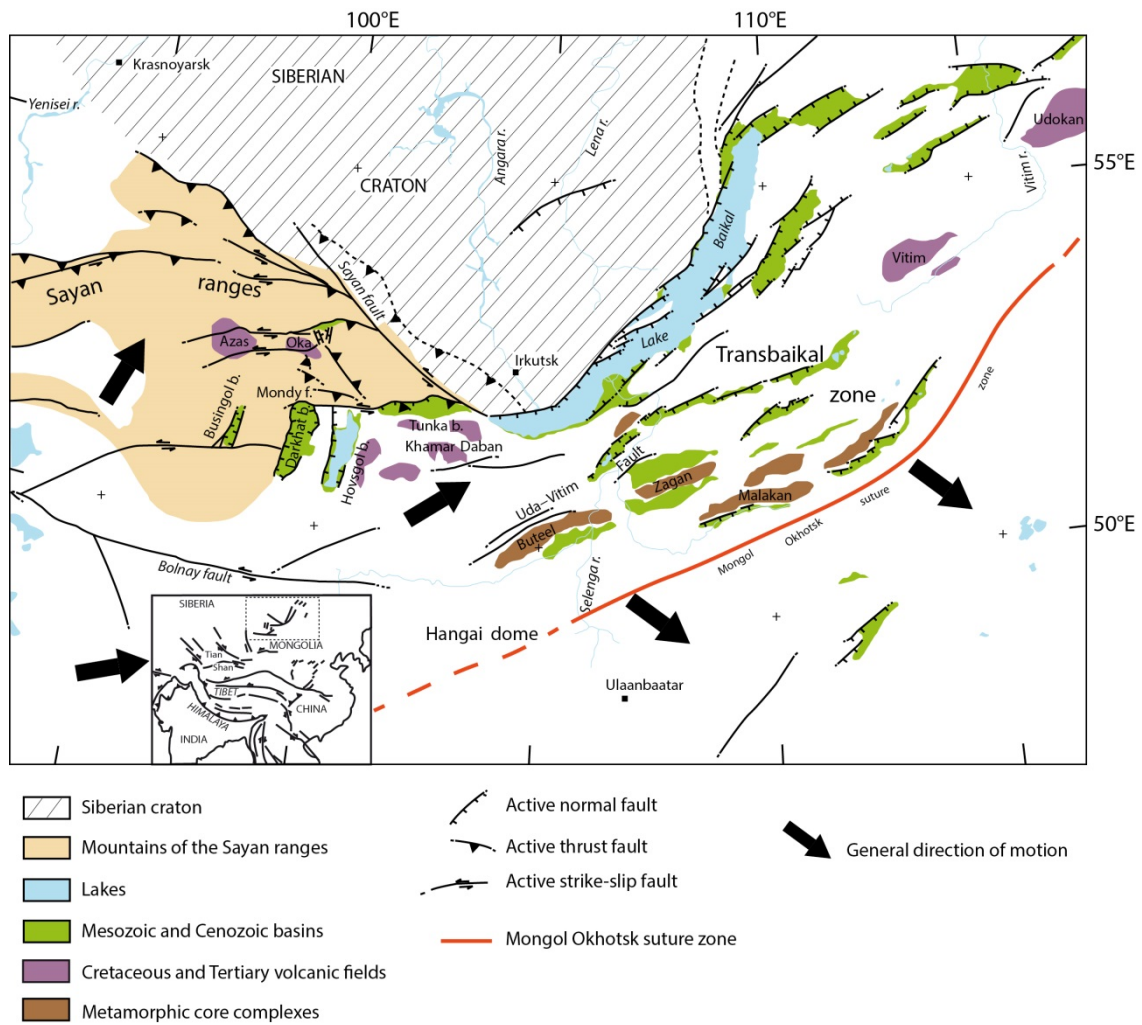


Figure 1.1.19 : Carte représentant la zone de suture supposée de l'Océan Mongol-Okhotsk (ligne rouge), en région Transbaïkale. Les dômes métamorphiques sont représentés en brun et les bassins en vert. Les flèches noires indiquent les mouvements actuels déduits des données GPS (Jolivet et al., 2013a).

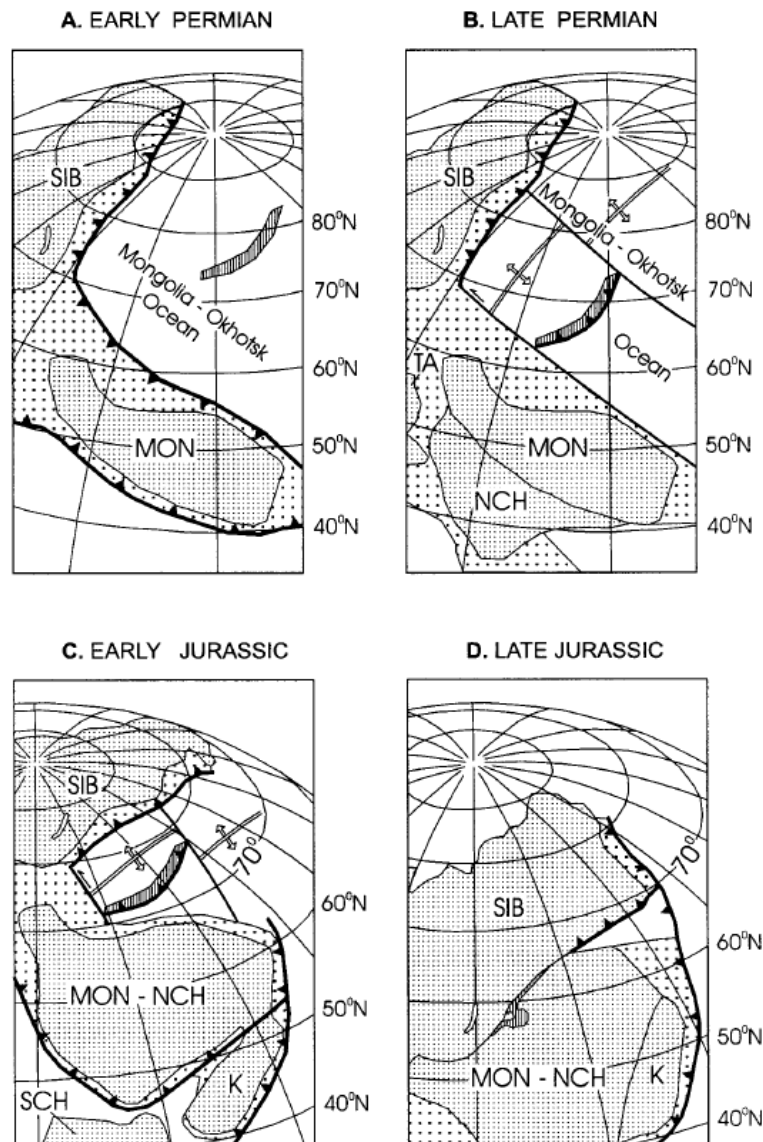


Figure 1.1.20 : Reconstructions paléogéodynamiques de la fermeture de l'Océan Mongol-Okhotsk, réalisées à partir de la synthèse des données de paléomagnétisme (Zorin (1999) et références associées). SIB = continent Sibérie ; MON-NCH = continent Mongolie-Chine du Nord ; MON = Bloc Mongolie ; NCH = Bloc Chine du Nord ; SCH = Bloc Chine du Sud ; TA = Bloc du Tarim ; K = Bloc de la Corée ; OIA = arc d'Onon

Les forts reliefs triasiques créés lors de l'orogénèse cimmérienne et les paléoreliefs du Paléozoïque supérieur continuent d'être érodés au Jurassique inférieur, poursuivant ainsi l'exhumation. L'ouest de la Mongolie (De Grave et al., 2002 ; Vassallo et al., 2007 ; Jolivet et al., 2007), le Gobi Altay (Vassallo et al., 2007 ; Jolivet et al., 2007) et le Tian Shan (Dumitru et al., 2001 ; De Grave et al., 2007, 2012 ; Jolivet et al., 2010), les Sayan (e.g. De Grave et Van den haute, 2002 ; Jolivet et al., 2013b) ainsi que le Nord Tibet sont concernés (e.g. Jolivet et al., 2001).

A partir du Jurassique moyen, une vaste surface de pénéplanation se forme en Asie centrale et va perdurer jusqu'au Tertiaire (cf. discussion et références dans Jolivet, in review). Les reliefs du Paléozoïques supérieur – Mésozoïque inférieur sont progressivement érodés au Jurassique et les données de thermochronologie basse-température enregistrent des taux d'érosion très faibles sur le long terme. Des restes préservés de cette surface ont été observés au Kazakhstan (e.g. Allen et al., 2001), au Tian Shan (e.g. Burbank et al., 1999 ; Dumitru et al., 2001 ; Jolivet et al., 2010), en Mongolie (e.g. Cunningham et al., 2003 ; Jolivet et al., 2007 ; Vassallo et al., 2007) et en Sibérie (De Grave et al., 2008 ; Jolivet et al., 2013b). Ces auteurs proposent des comparaisons et discussions détaillées de ces surfaces qui sont diachrones. Cependant, il est difficile de dater précisément leur formation.

De plus, dans de nombreuses régions d'Asie Centrale (Bassin Ouest-Sibérien, Bassin du Junggar, Bassin du Tarim, Bassin de Fergana, Bassin du Sichuan, Bassin du Qaidam, Bassin de l'Ordos, Vallée des lacs, région Transbaïkale, Mongolie Est), les sédiments déposés durant le Jurassique inférieur sont dominés par des dépôts de plaine alluviale, riches en épais horizons à charbon (e.g. Hendrix et al., 1992 ; Ritts & Biffi, 2001 ; Li et al., 2004 ; Le Heron et al., 2008). Ces dépôts généralisés sur l'ensemble de la zone confirment un régime tectonique relativement calme.

Dans la région du Tian Shan chinois, le régime tectonique au Jurassique inférieur et moyen ne semble pas être homogène. Dans le Bassin du Junggar par exemple, en raison de mouvements de rotation du bloc du Junggar, des évidences de compression et d'extension se superposent localement (e.g. Allen and Vincent, 1997 ; Liu et al., 2006). Les bords nord et ouest du bassin montrent des zones en compression, tandis que des bassins transtensifs se développent sur le bord sud du bassin, accompagnés d'une activité volcanique (Hou et al., 2009 ; Wang and Gao, 2012 ; Yang et al., 2013).

En Asie du Sud-Est, au Jurassique inférieur les granites post-orogéniques issus de la fusion des sédiments dans le prisme épaissi résultant de la collision entre les blocs de Qiangtang et Chine du Nord-Kunlun, se mettent en place dans la zone de Songpan-Garzê. Au sud, l'ouverture de l'océan Mésotéthys se poursuit, et le bloc de Lhassa remonte vers le nord (e.g. Metelkin, 2006, 2013). Jusqu'au Crétacé, l'Asie du Sud-Est est soumise à un régime tectonique majoritairement extensif, hormis au sud où le bloc de Lhassa commence à être accréte vers 140 Ma (e.g. Metcalfe, 2013).

Le régime extensif du Jurassique inférieur à moyen semble également s'exprimer par deux épisodes transtensifs dans la chaîne des Yanshan (à proximité de la faille de Tan-Lu) par du plutonisme et du volcanisme, ainsi que des failles normales et des grabens. Pour une discussion

détaillée sur ces épisodes prenant place dans le contexte de l'orogénèse Yanshan, le lecteur intéressé est invité à consulter les travaux de Faure et al. (2012).

1.1.3) Mésozoïque– Supérieur

1.1.3.1) Jurassique - Crétacé

L'activité tectonique au Jurassique supérieur est caractérisée par des mouvements extensifs dans la majeure partie de la bordure est de l'Asie, de la côte Pacifique jusqu'à la Sibérie. La formation de grabens et de dômes métamorphiques extensifs a lieu dans la région du Transbaikal et du Baikal Vitim (Fig. 1.1.19 et 1.1.23 ; e.g. Zheng et al., 1991 ; Zorin, 1999 ; Meng, 2003 ; Donskaya et al., 2008, 2013 ; Daoudene et al., 2009, 2013 ; Jolivet et al., 2009, 2013a ; Charles et al., 2010). Ces dômes métamorphiques extensifs se sont formés au Crétacé inférieur, entre 130 et 120 Ma (Donskaya et al., 2008 ; Daoudene et al., 2013 et les références associées). Ils sont généralement datés soit par des mylonites, soit par des coulées volcaniques intercalés dans les sédiments (e.g. Charles, 2010 ; Daoudene, 2011 et nombreuses références associées). De nombreux modèles tentent d'expliquer cette extension crétacée dans la partie est de la CAO. Le modèle le plus répandu est fondé sur une intense compression et un épaississement de la croûte (jusqu'à 60 km), provoqués par la collision entre les blocs Sibérie et Mongolie-Chine du Nord (Zheng et al., 1996, 1998 ; Darby et al., 2001a ; Lin et al., 2008 ; Donskaya et al., 2008). Cet épaississement est censé induire un effondrement orogénique rendu possible par le bord libre le long de la fosse Pacifique à l'est et entraînant l'extension Crétacé (e.g. Graham et al., 2001 ; Fan et al., 2003 ; Donskaya et al., 2008). D'autre part, la croûte épaissie devrait impliquer la présence d'une chaîne de montagnes de grande ampleur. Pourtant, aucune trace de métamorphisme ou de relief significatif datant du Jurassique supérieur ou du Crétacé inférieur et qui serait liée au sur-épaississement de la croûte n'a été observée dans la région Transbaïkale (i.e. directement au nord de la zone de suture) ou en Mongolie (e.g. Zorin, 1999 ; Jolivet et al., 2009, 2013a ; Jolivet, in review). Il est donc incertain que la chaîne du Mongol-Okhotsk ait existé (cf. 1.1.3.1 Jurassique - Crétacé). D'autres mécanismes ont été invoqués pour expliquer l'extension au Crétacé, parmi eux le retrait du panneau plongeant du Paléo-Pacifique (*slab rollback*) (Watson et al., 1987 ; Traynor and Sladen, 1995), la rupture du panneau plongeant de l'océan Mongol-Okhotsk (Meng, 2003) ou la délamination lithosphérique et une remontée du manteau (Wang et al., 2006). Un des modèles les plus récents est celui de Daoudene (2011), qui suggère une subduction à double vergence, du même type que la subduction ayant fermé l'océan Paléotéthys le

long de la suture de Jinsha (Roger et al., 2008, 2010). Une subduction de ce type conduit à une collision plus « modérée » entre les deux blocs et n'est donc pas associée à un épaissement crustal très fort, ni à du métamorphisme régional. Dans ce cas, le relief créé est probablement de petite taille et la chaîne de montagne n'a pas de racine métamorphique. De plus, Daoudene (2011) souligne le changement de direction de la plaque Izanagi au Crétacé, dont le déplacement devient quasiment sub-parallèle à l'orientation de la marge continentale est-asiatique. Il propose que l'extension qui affecte l'Asie orientale au Mésozoïque supérieur soit le résultat d'une diminution de la contrainte compressive vers 140 Ma, en bordure d'une lithosphère continentale anormalement chaude (Fig. 1.1.21).

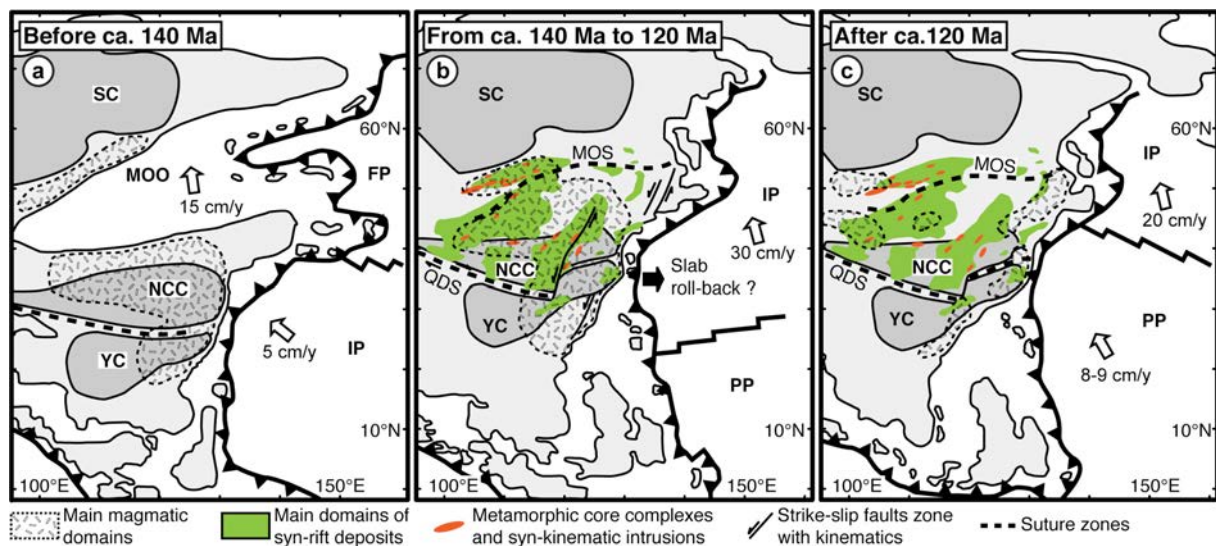


Figure 1.1.21 : Relations entre l'extension (nord-ouest/sud-est) dans le nord-est de l'Asie et les cinématiques dans le domaine paléo-Pacifique. a) contexte pré-extensif (MOO = Océan du Mongol-Okhotsk) – b) contexte syn-extensif – c) contexte post-extensif. NCC = Craton Chine du Nord, SC = Craton Sibérien, YC = Craton de Yangtze (ou Chine du Sud), MOS = Suture du Mongol-Okhotsk, QDS = zone de suture des Qinling-Dabie Shan, FP = Plaque de Farallon, IP = Plaque Izanagi, PP = plaque Pacifique (Daoudene, 2011, modifié d'après Maruyama et al., 1997).

Cette extension est également à l'origine de l'ouverture de nombreux bassins sédimentaires en Chine du nord-est au Jurassique supérieur. Les principaux bassins sont : Songliao, Hailar, Erlian, Gobi Est, Yingen et Sanjing (Fig. 1.1.22). Une forte activité magmatique a lieu en Chine du Nord à la même époque et dans les régions voisines (sud de la Mongolie, Péninsule Coréenne, Japon) et confirme le contexte extensif. L'expression maximale de l'extension est représentée par les dômes métamorphiques extensifs qui se forment au Crétacé inférieur et cités plus haut (e.g. Charles, 2010 ; Daoudene, 2011 et nombreuses références associées).

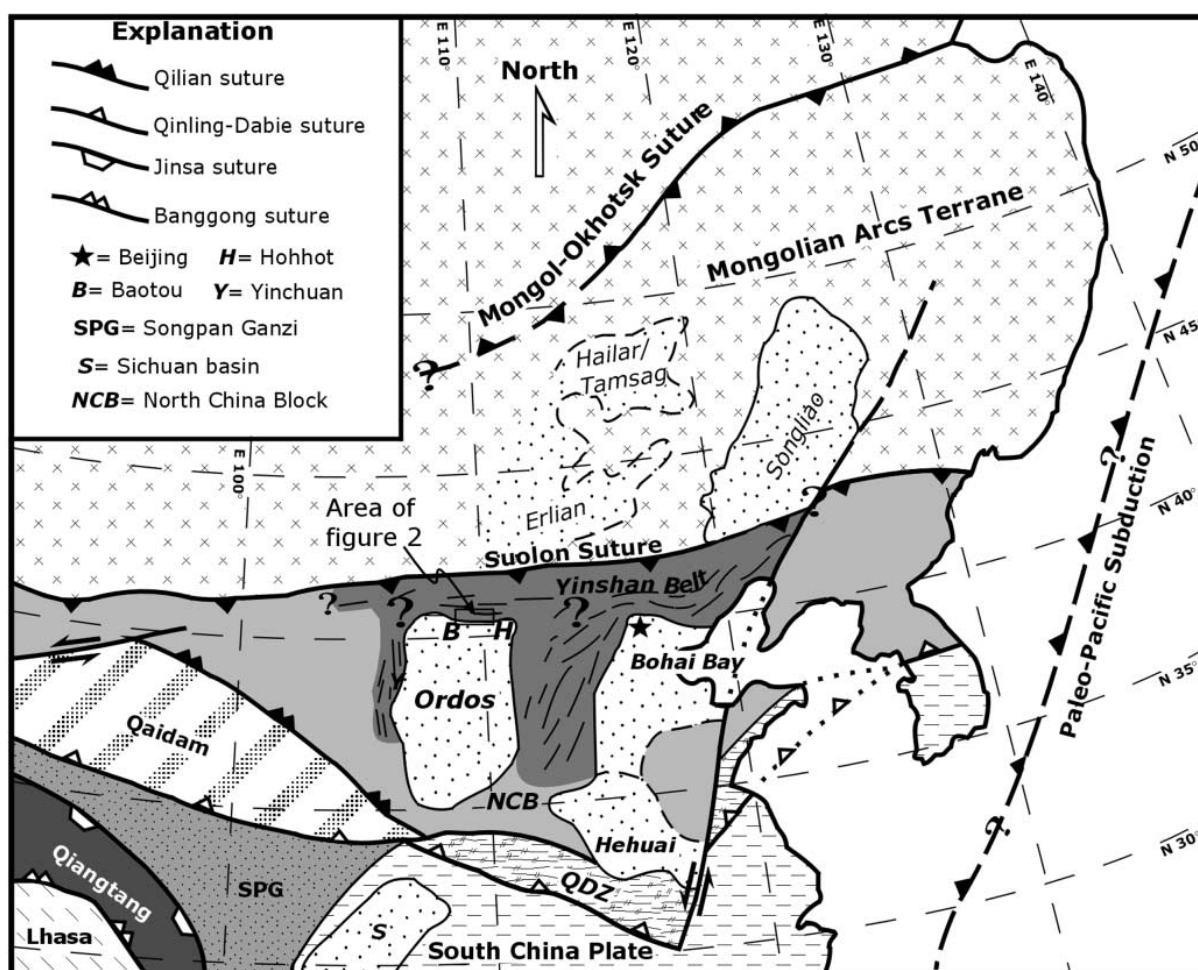


Figure 1.1.22 : Carte tectonique de l'Asie de l'Est indiquant les bassins méso-cénozoïques (fond blanc à points noirs) et les chaînes de montagnes méso-cénozoïques entourant le nord du Bassin de l'Ordos (gris sombre) (Ritts et al., 2001).

Dans le bassin de l'Ordos (Fig. 1.1.22, 1.1.23), le régime tectonique traduit plutôt un système compressif depuis la fin du Trias (Liu et al., 2013a), mise à part la présence de quelques failles normales sur la marge nord-ouest du bassin, durant le Trias supérieur (e.g. Liu, 1998; Ritts et al., 2004, 2009). Ce bassin reste donc sous un régime compressif, coincé entre l'orogène des Qinling-Dabie au sud, les Qilian Shan à l'ouest, les chaînes de Yinshan et Yanshan au nord et la chaîne de Taihangshan à l'est (Fig. 1.1.24, Faure et al., 2012 ; Liu et al., 2013a) et son régime tectonique diffère largement de celui du reste de l'Asie centrale au Jurassique et Crétacé, dominé par l'extension.

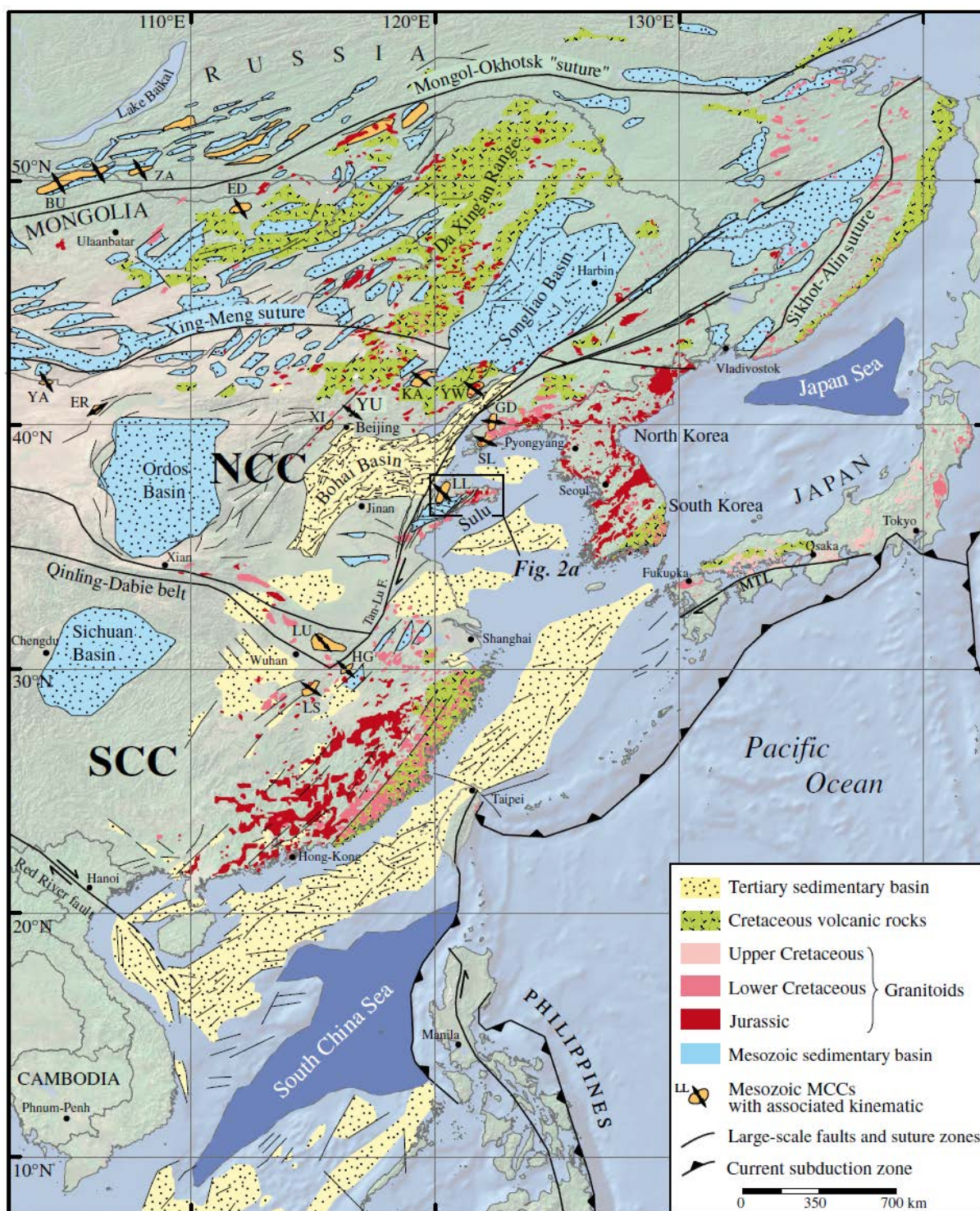


Figure 1.1.23 : Synthèse des cartes géologiques d'Asie de l'Est, où figurent les principaux marqueurs supposés de l'extension tardi-Mésozoïque à Cénozoïque : bassins sédimentaires, roches magmatiques, dômes extensifs et dômes métamorphiques extensifs. Abréviations des dômes et dômes métamorphiques extensifs : BU = Buteel-Burgutoy ; ED = Ereendavaa ; ER = Ertomiao ; GD = Gudaoling ; HG = Hongzhen ; KA = Kalaqin ; LL = Linglong ; LS = Lushan ; LU = Luotian ; SL = Sud Liaodong ; XI = Xishan ; YA = Yagan-Onch Hayrhan ; YW = Yiwulüshan ; ZA = Zagan. TLF = Faille de Tan-Lu ; NCC = Bloc de Chine du Nord ; SCC= Bloc de Chine du Sud (Charles et al., 2013 et références associées).

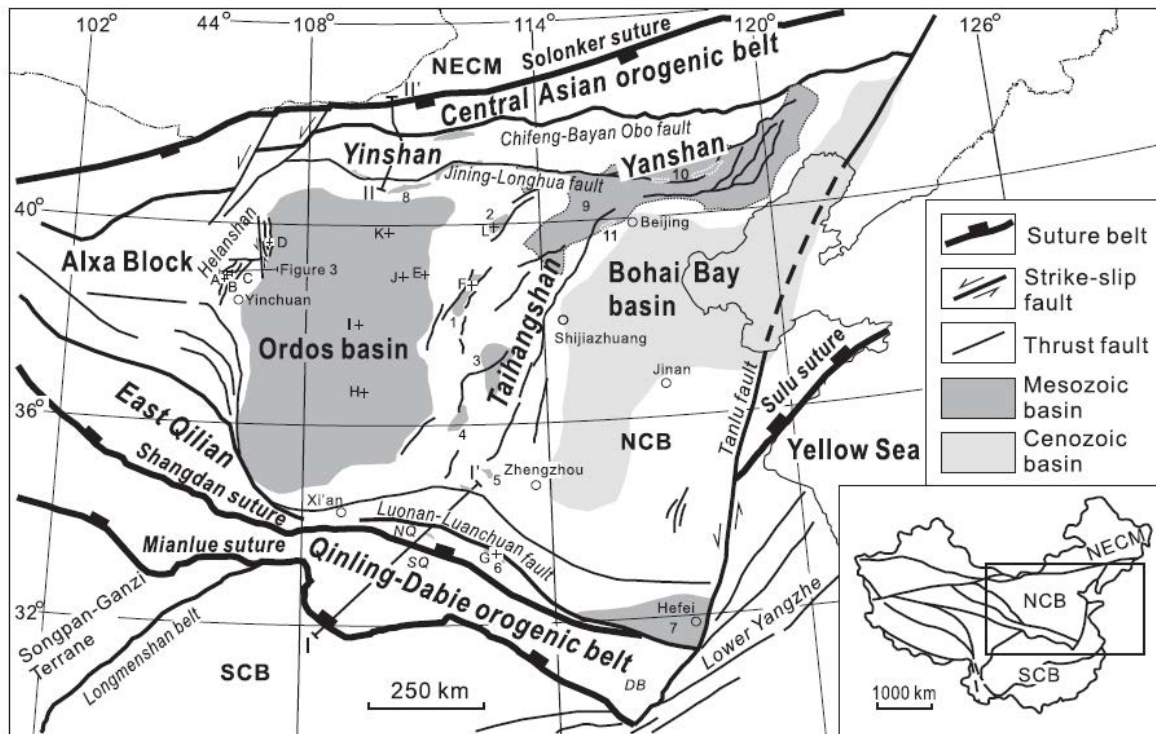


Figure 1.1.24 : Carte synthétique de l'Asie de l'Est indiquant la position des chaînes de Yinshan et Yanshan au nord du Bassin de l'Ordos (Liu et al., 2013a).

Dans le nord du Tibet, des marqueurs d'extension ont été observés, mais restent difficiles à interpréter. Sobel et al. (2001) décrivent les bassins dans l'Altun Shan comme étant en *pull-aparts*. Les bassins du Jurassique inférieur à moyen dans le Sud-Tarim et le N-Qaidam sont interprétés comme étant des bassins d'avant-pays en raison de la présence d'autres marqueurs indiquant le contexte compressif qui aurait suivi (Ritts & Biffi, 2000 ; Sobel et al., 2001 ; Wu et al., 2011). Pourtant, cette interprétation est contradictoire avec celle de Chen et al. (2003), basée sur des analyses tectoniques et sur l'architecture des sédiments, qui interprètent ces mêmes bassins comme des demi-grabens datant du Jurassique inférieur à moyen. A l'est et dans le sud du bassin du Tarim, le volcanisme alcalin et basaltique observé semble être une conséquence de l'extension (Zhang & Guo, 1998). Plus à l'est, dans la région des Qilian Shan et du Bassin de Jiuquan, des failles normales au Jurassique supérieur – Crétacé inférieur ont été identifiées sur les profils sismiques (e.g. Chen et al., 2003). Encore plus à l'est, les Bassins de Minle et Chaoshui dans le Corridor de Hexi (à l'est du Bassin de l'Ordos, Fig. 1.1.21) ont été interprétés comme étant des bassins en pull-apart formés en contexte extensif. Il est important de noter qu'aucune faille décrochante n'a pu être mise en évidence et que cette interprétation repose uniquement sur des arguments stratigraphiques (Vincent and Allen, 1999).

Cette extension est également observée dans la région du Tian Shan et plus loin à l'ouest où elle atteint la région de la Mer Caspienne, le domaine de Turan et celui du sud Kazakhstan (Fig. 1.1.6a ; e.g. Zonenshain et al., 1990 ; Nikishin et al., 1998b ; Allen et al., 2001 ; Brunet et al., 2003).

Au Jurassique supérieur, la région du Tian Shan continue de voir se former une vaste pénéplaine (cf. 1.2.4 Jurassique inférieur et moyen – Asie centrale), qu'on retrouve dans d'autres zones de la CAO et où elle n'est pas toujours synchrone (e.g. Dobretsov et al., 1996 ; Jolivet et al., 2007, 2009, 2010, 2013b ; Vassallo et al., 2007 ; De Grave et al., 2008).

A la transition entre le Jurassique et le Crétacé, la chaîne de l'Altay est réactivée possiblement du fait de l'orogénèse du Mongol-Okhotsk (e.g. De Grave et al., 2002, 2008). Toutefois la déformation liée à cette dernière n'est pas observée dans les Sayan Est et dans la région du Transbaïkal, qui sont plus près de la zone de suture (e.g. Van der Beek et al., 1996 ; Jolivet et al., 2009, 2013b). Il reste donc à trouver un autre mécanisme.

Une accélération de l'exhumation est également enregistrée par les données de thermochronologie basse température dans la région du Baikal (Van der Beek et al., 1996). Celle-ci n'apparaît pas dans la zone située plus à l'Ouest, dans les Sayan Est (Jolivet et al., 2013b).

1.1.3.2) Crétacé – Collision du Bloc de Lhassa

Alors que la majeure partie de l'Asie du Nord est soumise à un régime extensif, l'accrétion du bloc de Lhassa a lieu le long de la marge sud du continent, formant la suture de Bangong Nujiang (e.g. Allegre et al., 1984 ; Dewey et al., 1988). L'âge de cet événement est largement discuté. Les données varient en fonction de la méthode utilisée, allant de 171 ± 6 Ma (obtenus par datations U/Pb sur titanite) et issue de roches métamorphiques de bas grade provenant de la marge nord du bloc de Lhassa (Xu et al., 1985), à 118 ± 3 Ma (obtenus par datations U/Pb sur zircon) et provenant d'intrusions granitiques (Kapp et al., 2007). De récentes études (datations U/Pb sur zircons issus d'ophiolites) montrent que la zone était encore active à la fin du Crétacé inférieur. D'après Kapp et al. (2005, 2007), la marge nord du bloc de Lhassa aurait été subduite sous la marge sud du bloc de Qiangtang d'environ 90 km au Crétacé inférieur. Cette subduction continentale entraîne une obduction des roches de la suture de Bangong Nujiang vers le sud (Girardeau et al., 1984, 1985 ; Coward et al., 1988 ; Kapp et al., 2003a). A la fin du Crétacé inférieur et au Crétacé supérieur, le

retrait vers le sud du panneau plongeant de la Néo-Téthys aurait augmenté l'angle de subduction et entraîné la compression (Kapp et al., 2007 ; Leier et al., 2007) ou l'extension (Zhang et al., 2004).

En Mongolie orientale, la mise en place d'intrusions granitiques syn-cinématiques (138 – 125 Ma) et la formation d'un système de horsts et grabens accompagnent l'exhumation de dômes métamorphiques (e.g. Ereendava) qui indiquent le contexte extensif de cette région (cf. Daoudene, 2011). Les bassins sédimentaires en Chine du Nord-Est enregistrent une subsidence qui s'accélère au début du Crétacé pour continuer jusqu'à 100 Ma.

1.1.3.3) Transition Crétacé - Tertiaire

Enfin, à la transition Crétacé-Tertiaire (vers 65-60 Ma) plusieurs accrétions de blocs ont lieu sur la marge sud-ouest de l'Asie, telles que celle du Kohistan ou de l'arc du Dras, qui annoncent la collision avec l'Inde (e.g. Burg, 2011). Celles-ci sont possiblement responsables des réactivations enregistrées dans le Tian Shan et l'Altai par exemple (e.g. Jolivet et al., 2010 ; Glorie et al., 2010, 2012a). Elles seront suivies de nombreuses réactivations et surrections dues à la collision entre l'Inde et l'Asie, telles que la chaîne des Pamir, du Tian Shan, etc. (e.g. Avouac et al., 1993a ; Schwab et al., 2004).

1.1.4 Climat

Au Mésozoïque, la majeure partie des continents formant l'Asie du Nord actuelle est située à des latitudes moyennes allant de 30° à 60° dans l'hémisphère nord (mis à part le Bloc de Lhassa qui se trouve dans l'hémisphère sud). Les ensembles structuraux formant l'Asie du Sud actuelle se trouvaient quant à eux entre l'Equateur et le Tropique du Cancer (e.g. Metcalfe, 2006, 2013).

D'autre part, les limites Permien-Trias et Crétacé-Tertiaire sont bien connues pour les extinctions de masse des espèces (faune et flore) et sont également observées en Asie (REF e.g. Rees, 2002 ; Retallack, 2013). Certains auteurs les mettent en relation avec des changements climatiques majeurs qui auraient lieu aux mêmes périodes. Je ne détaillerai pas les possibles causes et les nombreuses théories qui tentent d'expliquer ces évènements.

Il est difficile de résumer les variations de climat à l'échelle de toute l'Asie durant le Mésozoïque, je me concentre donc sur l'Asie Centrale (cf. 1.2 La chaîne du Tian Shan).

1.2) La chaîne du Tian Shan

Avant de détailler la description de la chaîne et l'histoire de sa formation, il est intéressant de s'intéresser à sa structure lithosphérique. Celle-ci a pu être mise en évidence grâce à plusieurs méthodes géophysiques. Des données de tomographie sismique ont permis d'identifier une zone d'anomalie positive en-dessous du Tian Shan, qui plonge vers le nord (Fig. 1.2.1.a). Elle est présente sur plusieurs coupes transversales à la chaîne et peut être interprétée comme un domaine mantellique « froid ». Cette zone semble indiquer la présence d'un panneau de subduction plongeant vers le nord et peut être interprétée comme étant la subduction du bloc du Tarim sous le Tian Shan (Fig. 1.2.1.b ; e.g. Van der Voo et al., 1999). L'étude de Zhao et al. (2003), basée sur des données de sismologie et de gravimétrie, montre également une subduction du bloc du Tarim vers le nord. Le pendage de la subduction est de l'ordre de 35° (e.g. Zhao et al., 2003).

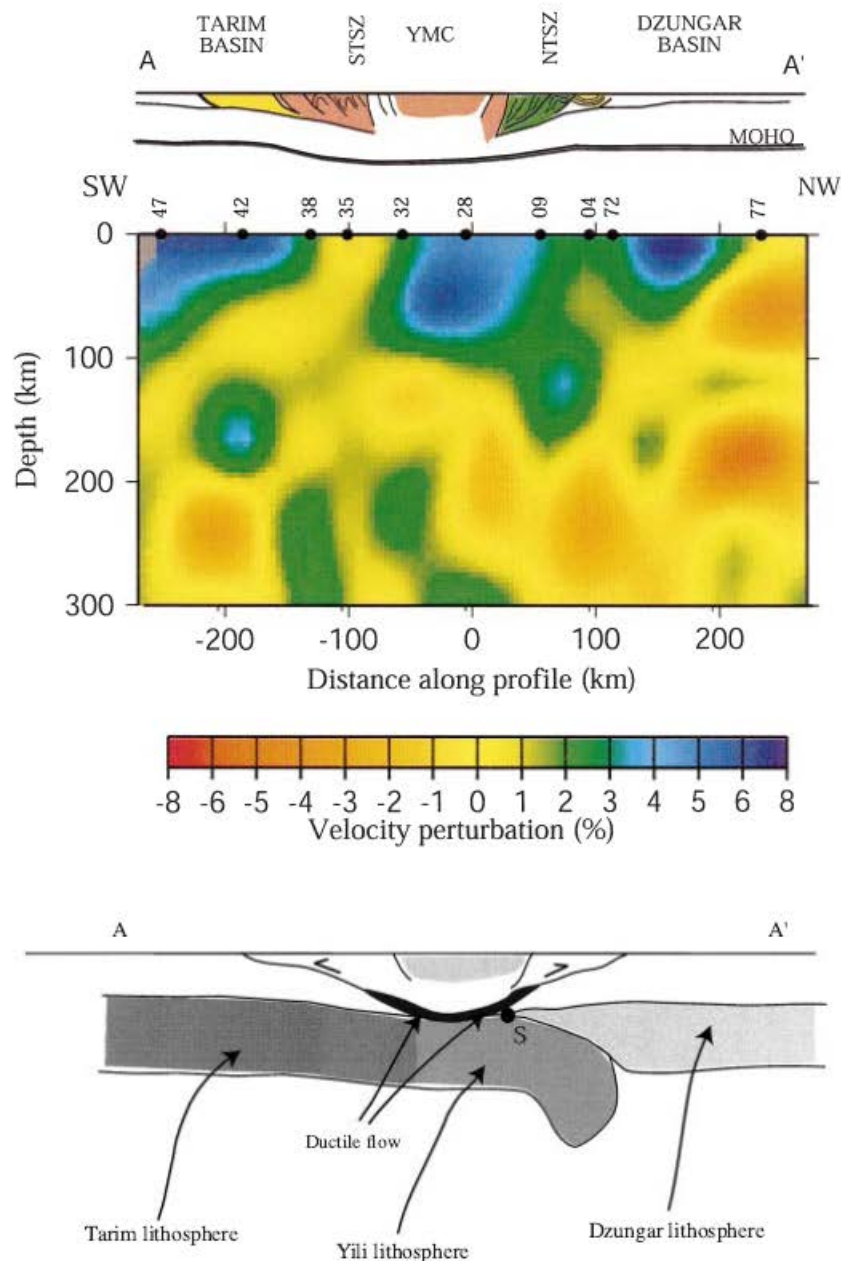


Figure 1.2.1 : Coupe tomographique et schéma d'interprétation. a) Coupe tomographique imageant la lithosphère et le manteau dans la chaîne du Tian Shan, le long d'un profil entre Karamay et Kuqa. Les vitesses P rapides indiquent une lithosphère plus épaisse (Poupinet et al., 2002). b) Schéma d'interprétation déduit des données géologiques et de la coupe tomographique (cf. Fig. 2) (Poupinet et al., 2002).

1.2.1 Introduction

1.2.1.1 La chaîne aujourd'hui

Le Tian Shan est actuellement une chaîne intracontinentale active située en Asie centrale, qui s'étend d'ouest en est sur plus de 2500 km, depuis le Kazakhstan et le Kirghizstan jusqu'à la Chine

(Fig. 1.2.2). Au nord-est, la chaîne est bordée par le Bassin du Junggar et au sud par le Bassin du Tarim. De nombreux bassins intramontagneux sont préservés, tels que les bassins de Chu (près de Bishkek, Kirghizistan), d'Ili (Almaty, Kazakhstan), de Naryn (Kirghizistan), d'Issyk-Kul (Kirghizistan) dans la partie ouest de la chaîne et les bassins de Yili (près de Yining, Chine), de Bayanbulak et de Turfan à l'est (Fig. 1.2.2).

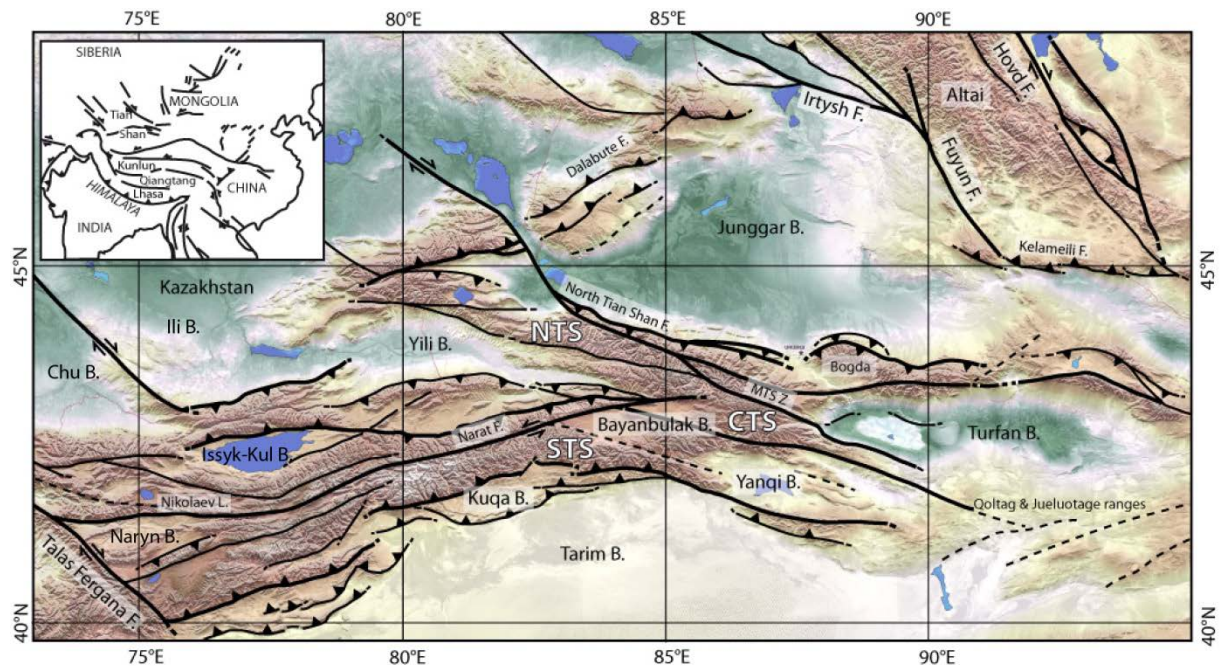


Figure 1.2.2 : Carte de la région du Tian Shan indiquant les principales unités composant la chaîne (NTS = Tian Shan Nord, CTS = Tian Shan Central, STS = Tian Shan Sud), les principales structures tectoniques, ainsi que les principaux bassins sédimentaires (Jolivet et al., 2013c).

1.2.1.2 Les bassins du Junggar et du Tarim

Le bassin du Junggar se trouve au nord de la chaîne du Tian Shan. De forme triangulaire, il couvre près de 130 000 km². Il est bordé par différentes chaînes de montagnes et les structures associées : les montagnes et la faille de Karamayi (ou Dalabute) au Nord-Ouest, l'Altay au nord-est, Kelameili à l'est et les failles « bordières » de la chaîne du Tian Shan (Fig. 1.2.3, 1.2.4) au sud. La nature du socle du bassin du Junggar est une véritable énigme. Certains auteurs y ont vu l'existence d'une croûte océanique paléozoïque (e.g. Hsü, 1988 ; Carroll, et al., 1990), d'autres une croûte continentale protérozoïque comme l'attestent les données de géophysiques (e.g. Gao, et al., 1998 ; Zhang, et al., 1984). En effet, l'épaisseur crustale a été estimée à environ 47 km (Li & Mooney, 1998 ; Liu, et al., 1990), ce qui semble suggérer la présence d'une croûte continentale. Six sous-bassins ont

été définis (Li, 1993), mais le plus important en terme de dépôts est la dépression située la plus au sud, le long du Tian Shan. Elle est la plus profonde et a recueillie environ 16 km de sédiments depuis le Permien (e.g. Qiu et al., 2005, 2008).

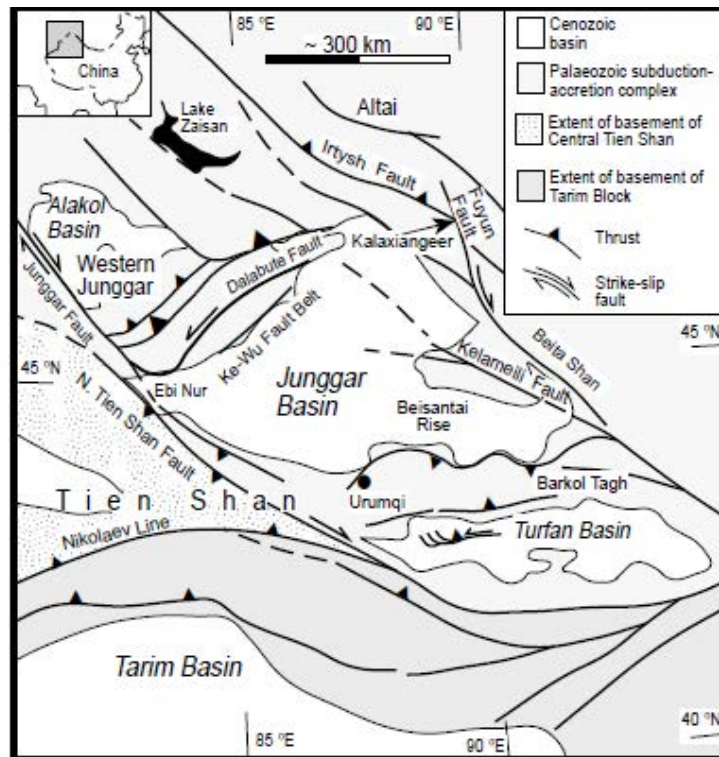


Figure 1.2.3 : Failles majeures autour du bassin du Junggar (Allen et al., 1997).

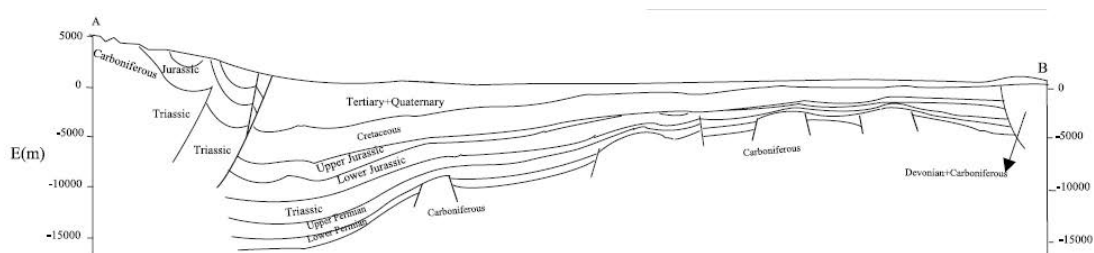


Figure 1.2.4 : Interprétation d'une ligne sismique orientée nord-sud à travers le bassin du Junggar et indiquant le dépôcentre du Sud-Junggar, où plusieurs milliers de mètres de sédiments ont été accumulés durant le Mésozoïque (Qiu et al., 2005, d'après Li, 1993).

Le bassin du Tarim se trouve au sud de la chaîne du Tian Shan. Il est de forme rhomboïdale et s'étend sur 560 000 km². Il est délimité à l'ouest par la chaîne des Pamirs et au Sud par la chaîne des Kunlun Shan, qui forment la bordure nord du plateau du Tibet (Fig. 1.2.5). Aujourd'hui, le bassin du Tarim est endoréique. D'après des études géophysiques, le socle du bassin du Tarim semble être

composé de gneiss archéens dans la zone centrale (Zhang, et al., 1984). Le socle affleurant sur les bordures du bassin est constitué de gneiss plissés de l'Archéen supérieur. Le Protérozoïque est composé de schistes, de méta-sédiments, de marbres et de dolomies (Allen, et al., 1991, 1993). Enfin, le Paléozoïque inférieur est déposé de manière discordante sur le Protérozoïque (Fig. 1.2.6).

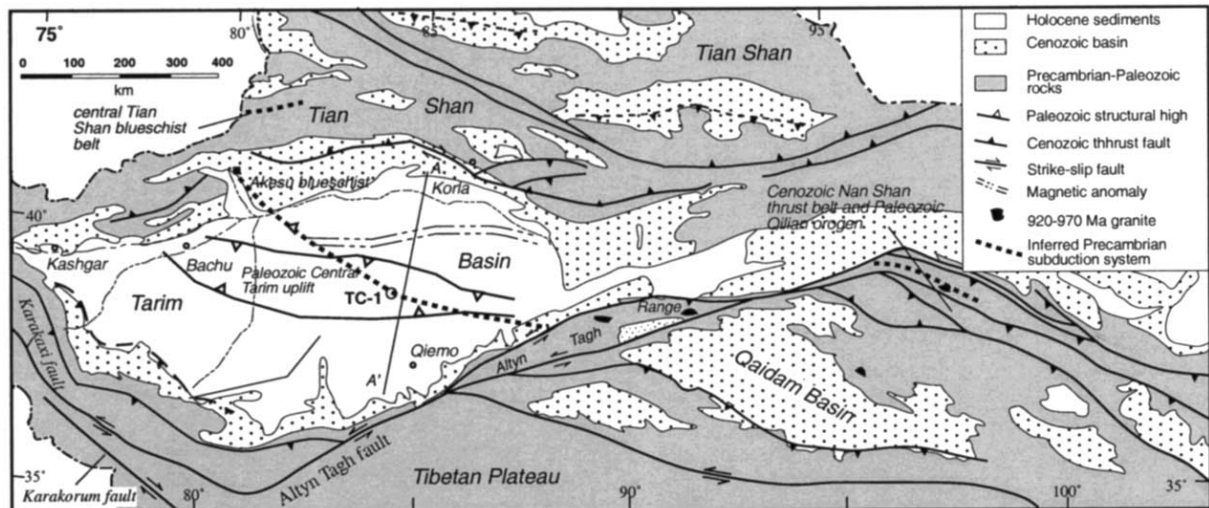


Figure 1.2.5 : Carte tectonique simplifiée du bassin du Tarim (Guo & al., 2005).

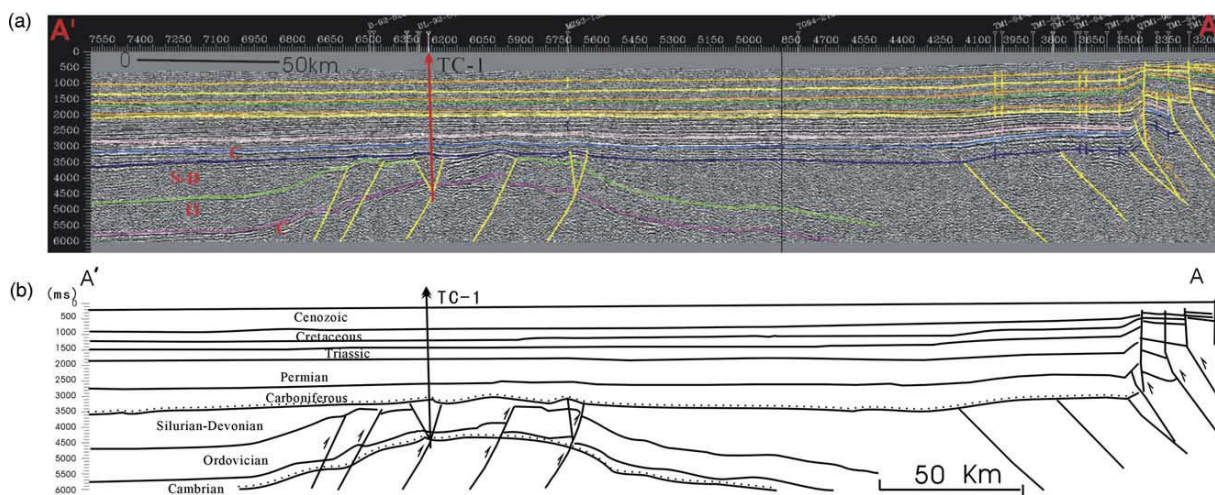


Figure 1.2.6 : Profil sismique composite à travers le bassin du Tarim (a), et son interprétation (b). L'exagération verticale est non-négligeable (cf. Figure 1.2.5 pour la position du profil) (Guo et al., 2005).

1.2.1.3 Les bassins intramontagneux (Yili, Bayanbulak et Turfan)

Les bassins de Yili et Bayanbulak se forment durant le Permien en raison de l'activité dextres sur les failles décrochantes au Permien (e.g. Wang et al., 2009 ; Jolivet et al., 2010).

Le bassin de Yili appartient au bloc de Yili, tandis que le bassin de Bayanbulak fait partie de l'unité Tian Shan Sud (STS) (Fig. 1.2.7). Le bassin de Yili (aussi appelé bassin de Yining, Fig. 1.2.8) est bordé au nord par le chaînon du Borohoro (appartenant au NTS) et au sud par le chaînon de la Narat. Le Borohoro est composé de deux chaînons (nord et sud), séparés par le bassin de Hexilagen, qui est de plus petite taille.

Le bassin de Bayanbulak est bordé par la faille de Narat au nord (Fig. 1.2.2), ainsi que de failles orientées NW-SE (leurs noms n'ont pas été déterminés) au nord-est et au sud-ouest du bassin.

Les séries mésozoïques affleurent peu et ces bassins intramontagneux ont été difficiles d'accès durant de longues années, ce qui explique le manque d'étude des séries sédimentaires.

Figure 1.2.7 : Carte géologique schématique du Tian Shan. 1. Faille du Nord Tian Shan – 2. *Main Tian Shan Shear Zone* – 3. Faille de Narat – 4. Faille de Baluntai – 5. Faille de Houxia – 6. Faille de Hongliuhe – 7. Faille de Sayram-Jinghe – 8. Ligne de Nikolaev – 9. Faille d'Atbashi-Inylchek (Charvet et al. 2011, mis en couleurs).

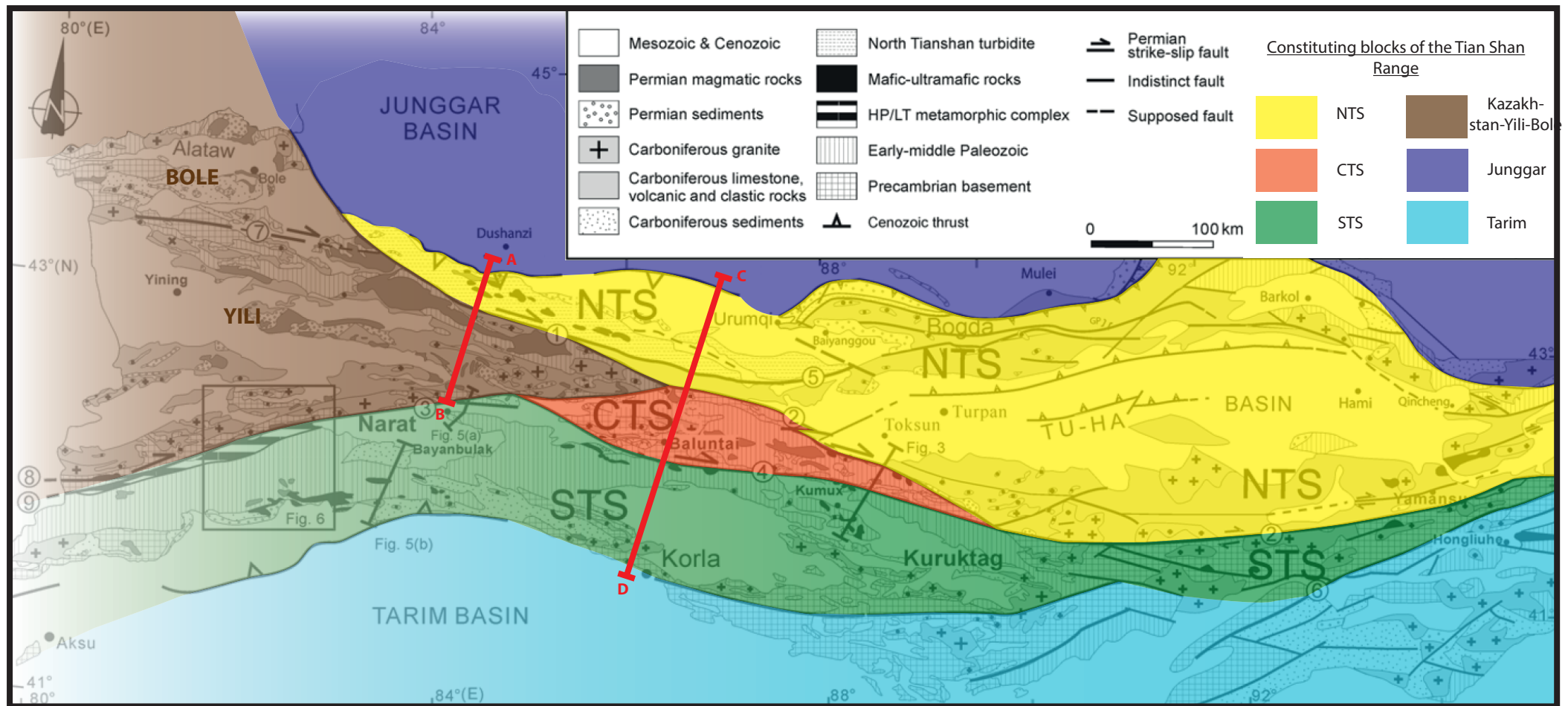


Figure 1.2.7

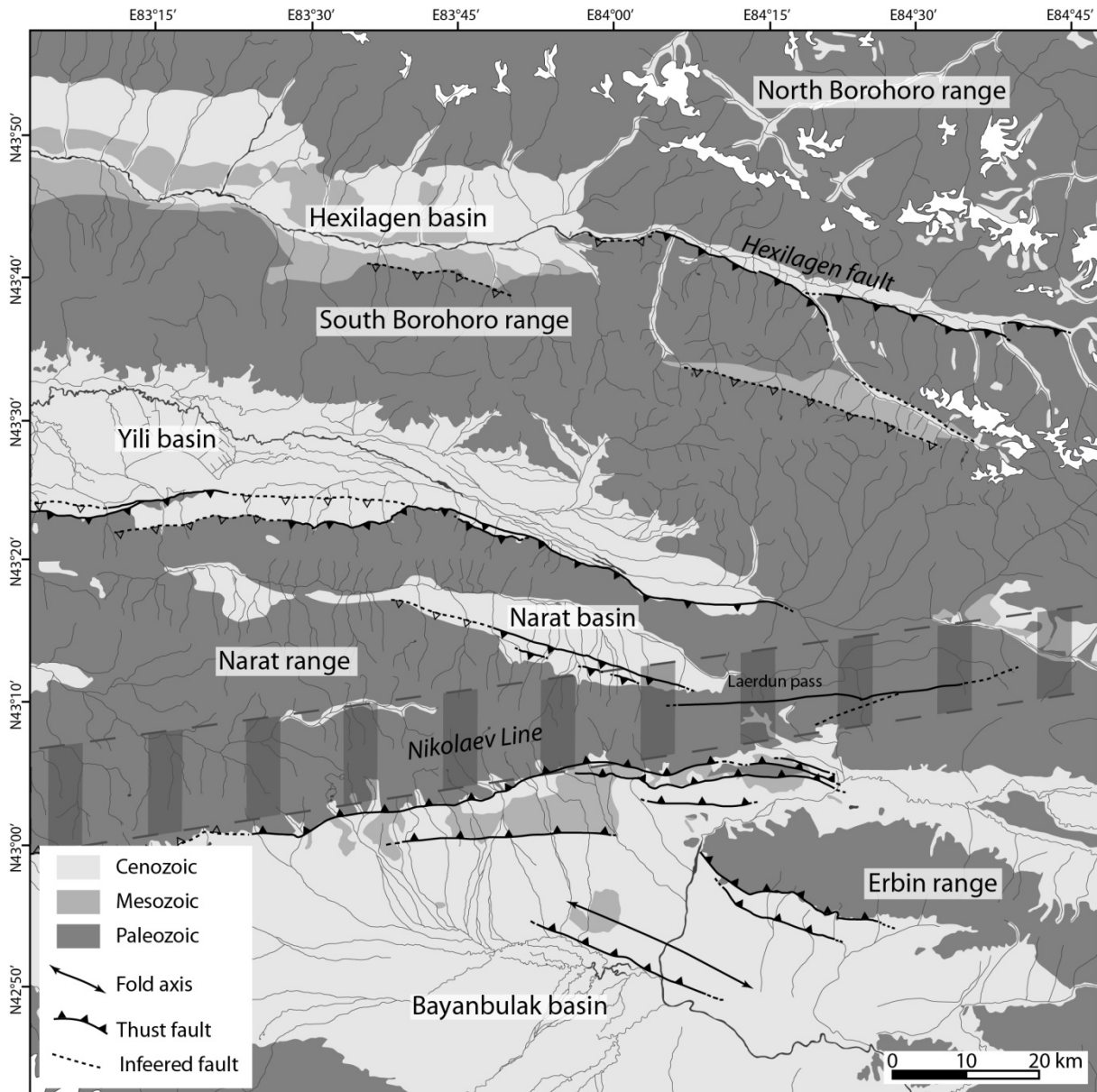


Figure 1.2.8 : Carte de la zone où sont situés les bassins intramontagneux de Bayanbulak, Narat, Yili (partie est) et Hexilagen, et principales structures tectoniques au cœur de la chaîne du Tian Shan (Jolivet et al., 2010).

Le bassin intramontagneux de Turfan est situé au nord-est de la chaîne du Tian Shan et au sud-est du bassin du Junggar. De forme allongée, il s'étend d'est en ouest en couvrant une superficie d'environ 50 000 km². Il est bordé par les branches nord et sud du Tian Shan Nord : les deux chaînes du Bogda Shan et du Barkol Tagh (ou Haerlike Shan) au nord et celle du Chöl Tagh (ou Jueluotage) au sud (Fig. 1.2.9). Le socle n'affleure que sur les bords du bassin. Il est composé de roches volcaniques, granitoïdes, volcano-sédimentaires et de dépôts marins profonds permien, formés dans le contexte de l'arc insulaire et du complexe d'accrétion du Tian Shan Nord (Allen et al., 1993a, 1995). Le dépoint mésozoïque du bassin de Turfan est situé dans la partie nord-ouest et est orienté

approximativement Est-Ouest. Environ quatre kilomètres de sédiments ont été déposés durant le Mésozoïque (Fig. 1.2.10 ; Allen et al., 1993a), et plus de sept kilomètres depuis le début de l'histoire du bassin (Shao et al., 1999).

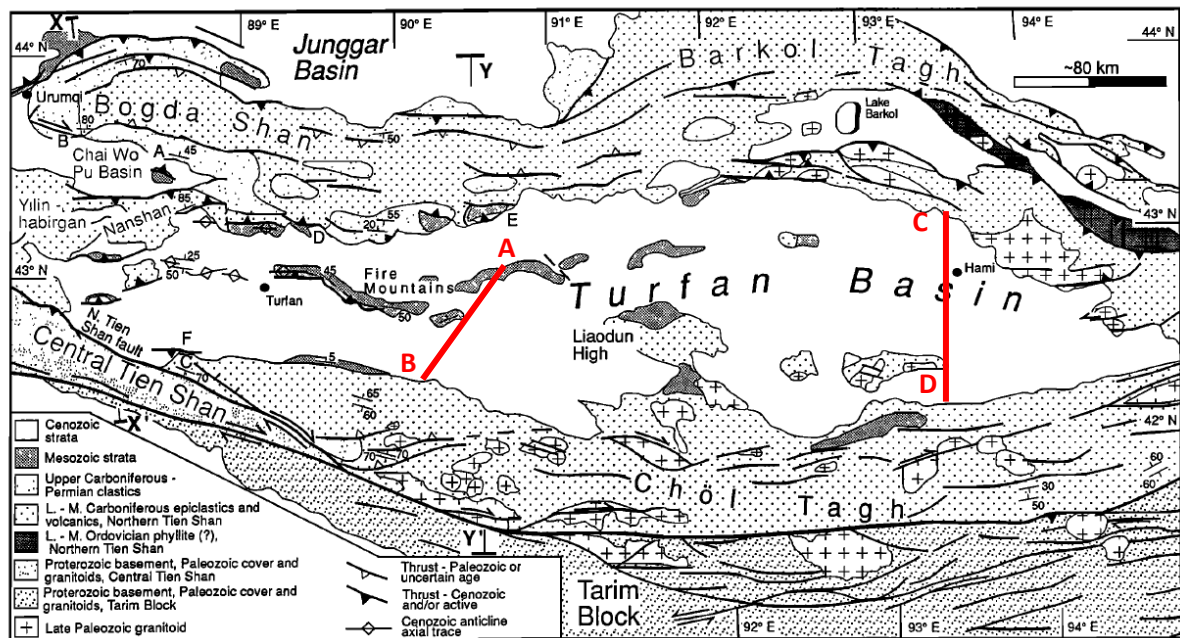


Figure 1.2.9 : Carte du bassin de Turfan et localisation des deux coupes (AB et CD) à travers le bassin (cf. Fig. 1.2.10) (Allen et al., 1993a).

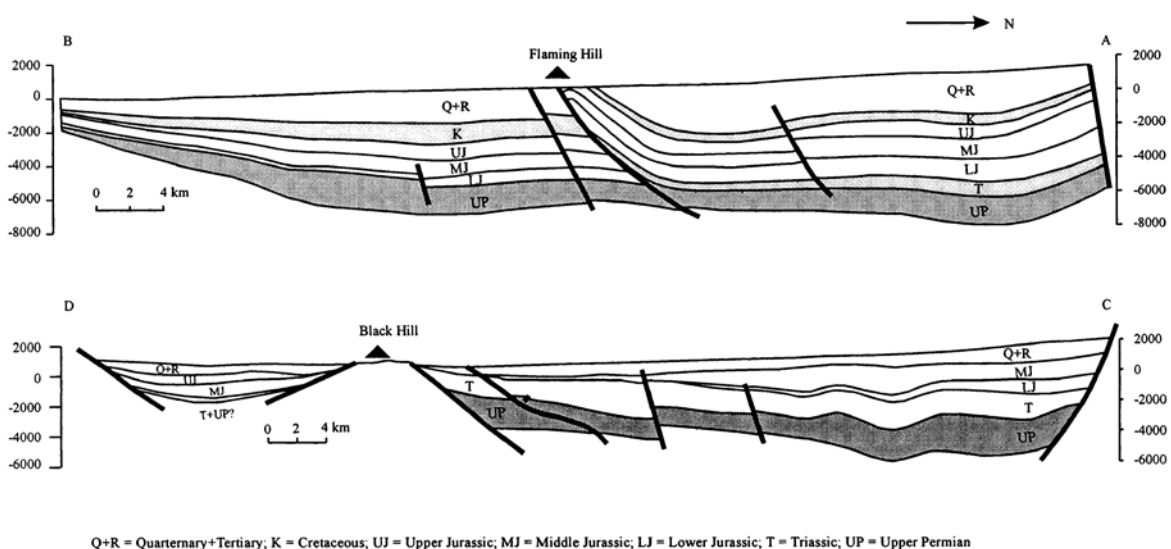


Figure 1.2.10 : Deux coupes (A-B et C-D) transverses au bassin de Turfan, issues de l'interprétation de lignes sismiques (voir leur position en rouge sur la Figure 1.2.9) (Shao et al., 1999). La coupe A-B indique l'emplacement où les séries mésozoïques affleurent localement, au milieu du bassin (*Flaming Mountains*), en raison d'un chevauchement.

Au nord-est du bassin de Turfan, le bassin de Chaiwopu est d'ampleur plus modeste (superficie d'environ 3 000 km²) et constitue également un dépocentre durant le Mésozoïque. Il est séparé du bassin de Turfan par le chaînon du Nanshan (Fig. 1.2.9).

Au nord de ce dernier bassin, se trouve le dépocentre mésozoïque de Houxia, situé aujourd'hui dans une petite vallée (superficie d'environ 1 500 km²) orientée Est-Ouest et localisée non loin de la vallée de la rivière Urumqi, au Sud de la ville du même nom. Les sédiments jurassiques (Guo et al., 2005) y affleurent et reposent de manière discordante sur les roches métamorphiques carbonifères (Chen, 2010).

1.2.2 L'histoire paléozoïque moyen : Structuration de la lithosphère et formation de la chaîne

1.2.2.1 Les principaux événements tectoniques

La chaîne du Tian Shan s'est formée au cours du Paléozoïque suite à une histoire complexe de subduction – collision – accrétion. De nombreux modèles ont proposé une interprétation de la fermeture des océans paléozoïques et de la formation de la CAO (cf. 1.1.1 Paléozoïque supérieur – Mésozoïque inférieur), dont fait partie la chaîne du Tian Shan. Ainsi, les théories développées au cours des années 1990 par Sengör (Sengör et al., 1993 ; Sengör & Natal'in, 1996 ; etc.) imaginent l'Asie Centrale comme résultant de l'accrétion de nombreux arcs et blocs continentaux formant un « super-collage », l'Asie Centrale actuelle. Ce modèle est insuffisamment contraint par des données géologiques ou géochronologiques. D'autres modèles ont été proposés (e.g. Laurent-Charvet, 2001 ; Charvet et al., 2011), et depuis les études de Windley et al. (1990) et Gao et al. (1998), il est admis que l'histoire paléozoïque du Tian Shan résulte de deux principales phases : l'une au Paléozoïque moyen et l'autre au Paléozoïque supérieur. Il est également généralement admis que la construction de la lithosphère du Tian Shan est achevée à la fin du Carbonifère - début du Permien.

L'histoire de la chaîne débute à l'Ordovicien avec la subduction de la croûte océanique de l'océan du Tian Shan central (*CTS ocean*) sous la marge active du Tarim (Fig. 1.2.11). Elle entraîne la formation de l'arc volcanique du Tian Shan central (*CTS arc*) sur la marge nord du Tarim. Suite au rifting silurien qui affecte la marge nord du Tarim, l'arc du CTS se détache de la marge. L'autre conséquence de cette phase d'extension est l'ouverture du bassin d'arrière-arc du Tian Shan sud (*STS back-arc basin*), qui a lieu au Silurien supérieur – Dévonien (e.g. Charvet et al., 2011).

La première phase majeure de formation du Tian Shan est marquée par deux évènements. Dans un premier temps, la **fermeture de l'océan du Tian Shan central** au Dévonien moyen à supérieur entraîne la collision entre l'arc du CTS au Sud et l'ensemble Kazakhstan-Yili au Nord (Fig. 1.2.11, e.g. Charvet et al., 2011). Dans un deuxième temps, la **fermeture du bassin d'arrière-arc du STS** a lieu au Dévonien supérieur – Carbonifère inférieur d'après les dernières études de Charvet et al. (2011). Cette interprétation est basée sur l'activité d'un arc magmatique dans la plaque chevauchante (Tarim) à la transition Dévonien-Carbonifère. Pourtant, cet épisode reste très débattu en raison de la nature controversée de la vergence de la subduction de l'océan Tian Shan sud (e.g. Laurent-Charvet, 2001). Certains auteurs argumentent en faveur d'une subduction à vergence nord, en s'appuyant sur le magmatisme d'arc continental observé au Nord de la suture, daté par Gao et al. (2009a) à 460-330 Ma, la présence d'unités ophiolitiques, de roches de faciès schistes bleus au nord et par l'absence d'un arc calco-alcalin au sud (e.g. Burtman, 1975, 1980 ; Gao et al., 1995 ; Chen et al., 1999 ; Xiao et al., 2004d ; Zhang et al., 2007 ; Gao et al., 2009a ; Alexeiev et al., 2011 ; Xiao et al., 2013 et références associées). La subduction commencerait donc bien plus tôt que dans le modèle de Charvet et al. (2011). Gao et al. (2009a) suggèrent l'Ordovicien supérieur (460Ma) ou le Silurien moyen. D'autres auteurs observent sur l'actuelle marge nord du Tarim, des basaltes d'arc insulaire (OIB) et de type dorsales océaniques (MORB) datant du Dévonien moyen – Carbonifère inférieur, et argumentent en faveur d'une subduction à vergence sud (e.g. Shu et al., 2002 ; Charvet et al., 2007, 2011 ; Wang et al., 2011). Le lecteur trouvera une discussion plus détaillée et plus argumentée concernant les conditions de subduction et de fermeture de l'océan du Tian Shan Sud dans la récente synthèse de Xiao et al. (2013). Au Dévonien supérieur, une nouvelle zone de subduction se forme au nord de l'ensemble Kazakhstan-Yili –CTS, où l'océan Tian Shan Nord est subduit vers le sud. L'arc volcanique du Tian Shan Nord se forme, ainsi qu'un prisme d'accrétion (e.g. Charvet et al., 2011). Au Carbonifère inférieur, la fermeture de l'océan Tian Shan Sud entraîne l'accrétion de l'ensemble arc du Tian Shan Central –Kazakhstan - Yili au nord et du craton du Tarim au sud (e.g. Gao et al., 2009a ; Wang et al., 2011).

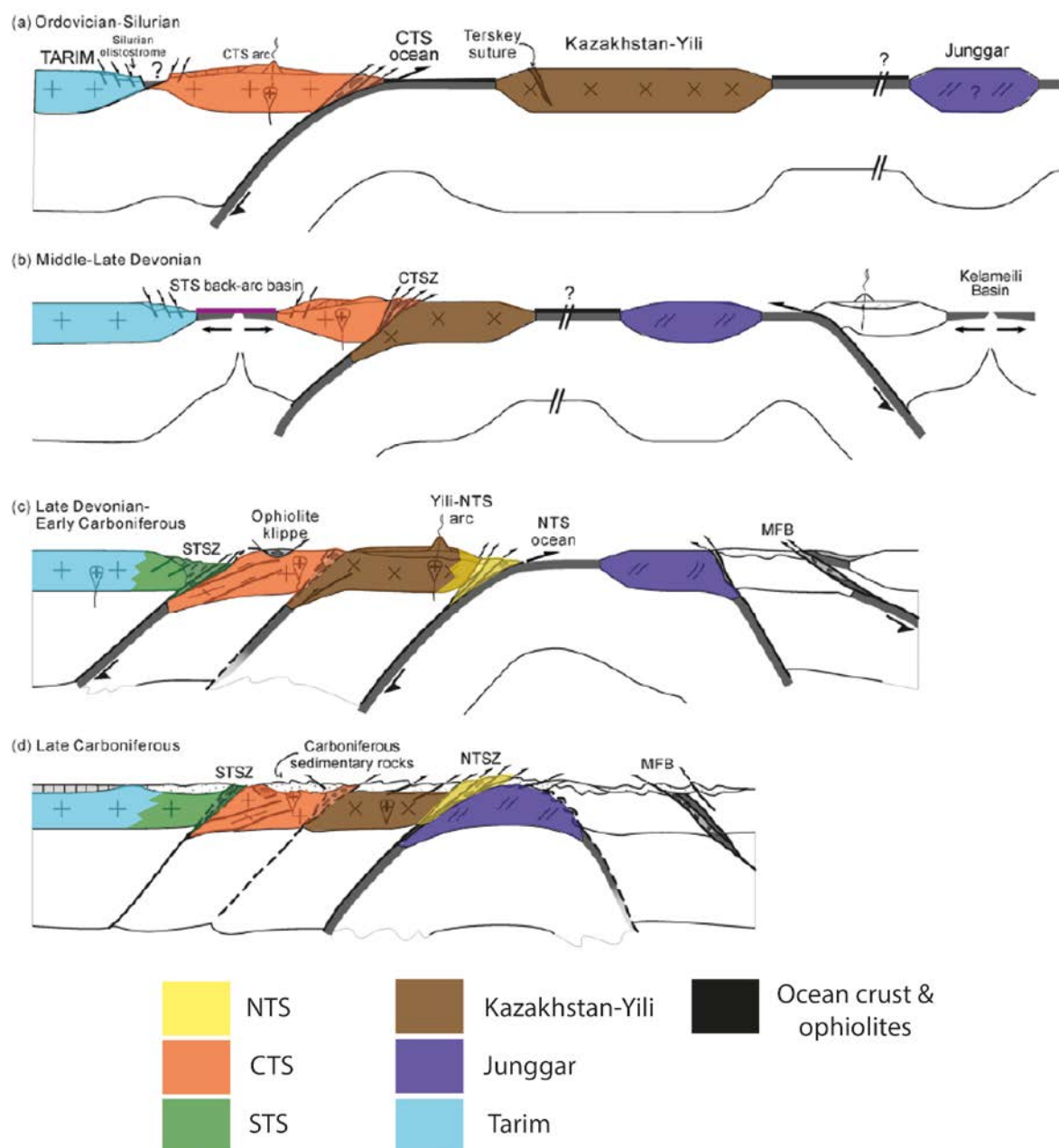


Figure 1.2.11 : Modèle géodynamique résumant l'accrétion des blocs qui forment la chaîne du Tian Shan chinois. Pour des raisons de simplifications, l'arc du NTS est dessiné côte à côte du Bloc de Yili, alors qu'il se forme dans le Bloc de Yili (Charvet et al., 2011, mis en couleurs).

La seconde phase majeure de formation du Tian Shan consiste en la fermeture du paléo-océan du Tian Shan Nord et la collision entre les blocs de CTS-Kazakhstan-Yili, le prisme d'accrétion du NTS et le bloc du Junggar au Carbonifère (e.g. Wang et al., 2009 ; Charvet et al., 2011). Celle-ci fait l'objet d'une longue controverse. Deux modèles tentent d'expliquer l'accrétion finale qui a mené à la formation de la chaîne. Le premier modèle est basé sur les laves datant du Carbonifère supérieur – Permien inférieur dans le bloc de CTS-Kazakhstan-Yili, qui indiquent une mise en place dans un rift

continental (e.g. Xiao et al., 1992 ; Che et al., 1996 ; Xia et al., 2004, 2008, 2012). Les données géochimiques et les datations U/Pb sur zircons issues de roches du Carbonifère inférieur à moyen, ont démontré que le domaine océanique du Tian Shan Nord a dû exister pendant une durée relativement courte au Carbonifère inférieur à moyen, et la fermeture du domaine entraîne la formation d'unités ophiolitiques datées du Carbonifère moyen. Suite à la fermeture, un régime extensif post-orogénique se met en place au Carbonifère supérieur et au Permien, et permet la mise en place d'un volcanisme basique typique d'une zone de rift dans le bloc de Yili, mais aussi dans les bassins environnants (Xia et al. 2004). Depuis, ce modèle évolue sans cesse, puisque l'interprétation de ces laves évolue, ainsi que la quantité et la précision des données les concernant. Wang et al. (2009) montrent que le magmatisme est continu à partir du Carbonifère inférieur jusqu'au Permien. La composition de ces magmas évolue au cours du temps, allant d'un magma calco-alcalin occasionnellement transitionnel durant le Carbonifère, à des magmas calco-alcalins, transitionnels et alcalins co-existant durant le Permien. Ainsi, les laves carbonifères seraient plutôt interprétées comme mises en place durant la subduction vers le Sud de l'Océan du Tian Shan nord sous le bloc de Yili (e.g. Wang et al., 2006, 2009 ; Charvet et al., 2011). Cette subduction prend fin avec la fermeture de l'océan Tian Shan Nord au Carbonifère supérieur-Permien inférieur et la collision entre les blocs de Yili et du Junggar (Fig. 1.2.12, 1.2.13, 1.2.14, e.g. Gao et al., 1998 ; Wang et al., 2006 ; Charvet et al., 2007 ; Shu et al., 2010 ; Charvet et al., 2011). Les magmas permien se sont donc mis en place dans un contexte post-collisionnel. La signature mantellique de ces roches peut être expliquée soit par la rupture du panneau plongeant, soit par délamination lithosphérique (e.g. Han et al., 2009 ; Chen and Jahn, 2004 ; Shu et al., 2010). Ces roches du Carbonifère-Permien ont été datées par analyses U/Pb sur zircon ou analyses Ar/Ar sur roches totales (cf. synthèse dans Wang et al., 2009).

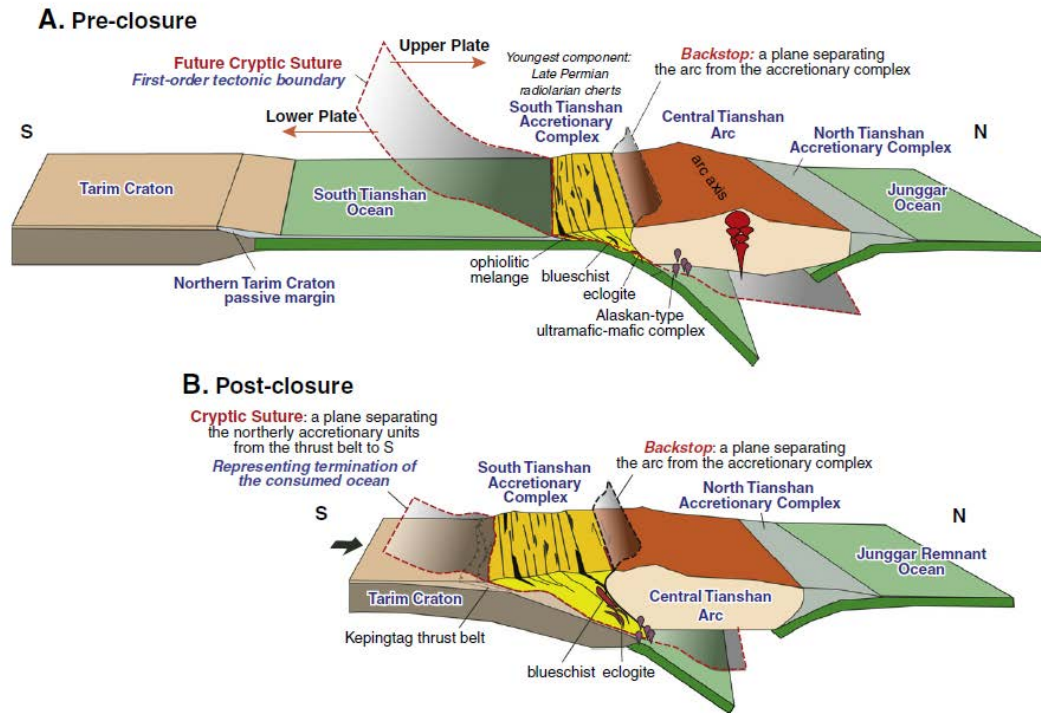


Figure 1.2.12 : Modèle géodynamique expliquant la fermeture de l'Océan du Tian Shan Sud, du Carbonifère supérieur au Permien. (A) Avant la fermeture et (B) après la fermeture de l'Océan du Tian Shan Sud (Xiao et al., 2013).

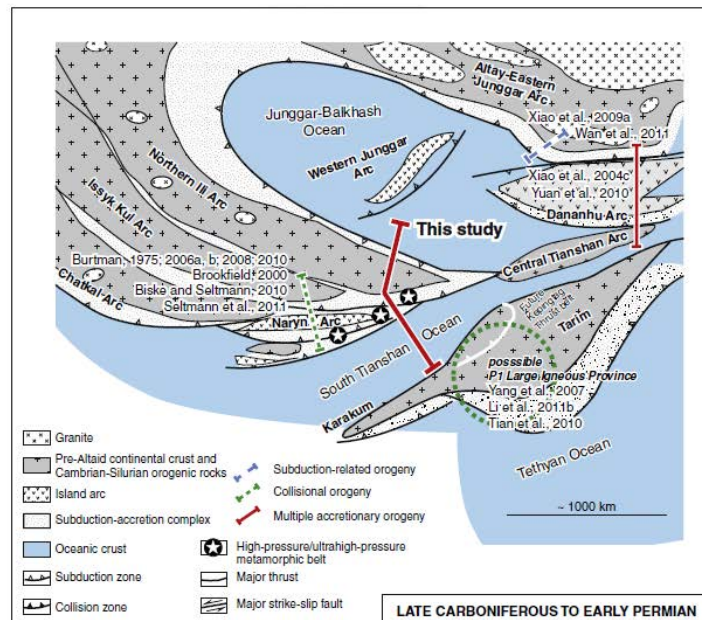


Figure 1.2.13 : Reconstruction palinspastique et tectonique de la fermeture de l'océan du Tian Shan Sud durant le Carbonifère supérieur – Permien (Xiao et al., 2013).

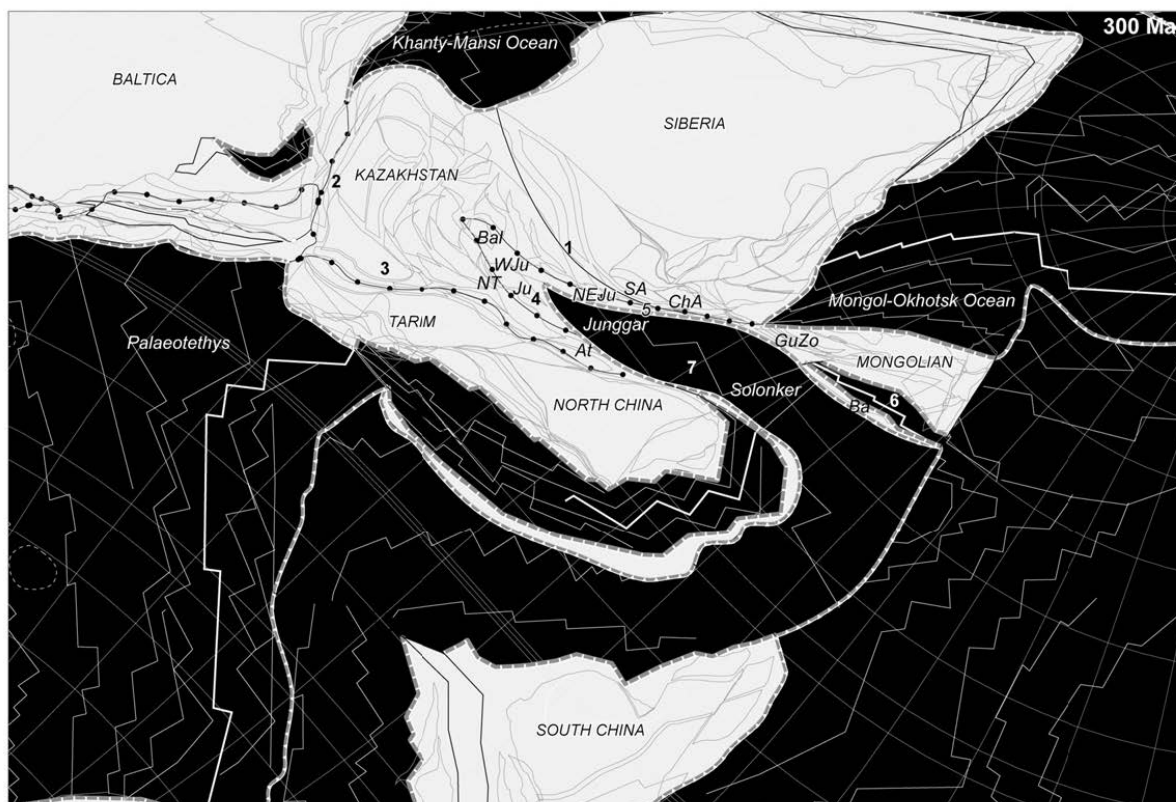


Figure 1.2.14 : Reconstruction de la position des plaques tectoniques au Carbonifère supérieur – Permien inférieur, indiquant les principaux océans paléo-mésozoïques qui se ferment durant le Mésozoïque. Bal = Balkash, JuB = Junggar Basin, NT = North Tianshan, WJu = Western Junggar, NEJu = northeastern Junggar arc, ChA = Chinese Altai, Ba = Baolidao, At = Atasbogd, GuZo = Gurvansayhan-Zoolen (Wilhem et al., 2012).

1.2.2.2 Le partitionnement de la chaîne du Tian Shan

La structure lithosphérique du Tian Shan résulte donc de l'accrétion de blocs continentaux et d'arcs volcaniques durant le Paléozoïque.

Au nord, le domaine du NTS est composé de dépôts paléozoïques supérieurs (flyschs du Carbonifère moyen ; séries volcaniques, tuffs et carbonates du Carbonifère inférieur). En revanche, aucune roche du Paléozoïque inférieur n'a été observée. Le NTS comporte deux branches : la branche nord est principalement constituée de l'arc carbonifère de Bogda-Haerlike (fruit de la subduction entre le NTS et le Junggar) (e.g. Charvet et al., 2011) et la branche sud est continue sur l'ensemble de la chaîne, à l'est comme à l'ouest (jusqu'au Kazakhstan) (Fig. 1.2.11). Cette branche est constituée de différentes sous-unités, séparées par des décrochements, et composées entre autres de volcanites carbonifères, probablement formées dans le contexte d'un arc insulaire mature ou d'une marge active continentale (Allen et al., 1993b). Le NTS est bordé au nord par de grands accidents cénozoïques, qui font chevaucher la chaîne du Tian Shan sur le bassin du Junggar (Allen et

al., 1995 ; Avouac et al., 1993a). Le Tian Shan nord (NTS) est séparé du bloc de Yili à l'Ouest par la faille du Nord Tian Shan (*North Tian Shan fault*), du Tian Shan central (CTS) et du Tian Shan sud (STS) respectivement au centre et à l'Est par la *Main Tian Shan Shear Zone* (MTSZ) (Fig. 1.2.11 ; e.g. Charvet et al., 2011).

Le CTS est composé d'un socle protérozoïque métamorphique, considéré comme identique à celui du Tarim (Gao et al., 1998) (gneiss, amphibolites, etc. ; Gao et al., 2009a ; XBGMR, 1993 ; Hu et al., 1998 ; Hu et al., 2000 ; Liu et al., 2004 ; Zhu & Song, 2006 ; Li et al., 2007 ; Yang et al., 2007), qui comporte de nombreuses intrusions granitiques datant de l'Ordovicien-Silurien-Devonien (490-380 Ma), ainsi que des sédiments et des roches volcaniques datant de l'Ordovicien-Silurien (Yang et al., 2007 ; Li et al., 2008). D'après les données géochimiques, il s'agit d'une ancienne marge active continentale (Laurent-Charvet, 2001 ; Wang, 2006 ; Wang et al., 2006 ; Wang et al., 2007b ; Zhu et al., 2005a). Au Permien supérieur, des molasses continentales non-déformées sont déposées en discordance sur les séries paléozoïques métamorphisées. Le CTS s'arrête latéralement vers l'Ouest, où il est bordé par le bloc de Yili (Fig. 1.2.7, 1.2.15, 1.2.16). Celui-ci est composé d'un socle protérozoïque, recouvert de sédiments marins du Paléozoïque inférieur (Wang et al., 2006 ; Gao et al., 1998 ; XBGMR, 1993 ; Chen et al., 1999). Au Nord du bloc de Yili, séparé par la faille de Sayram-Jinghe se trouve le microbloc de Bole, qui semble avoir un enregistrement sédimentaire assez différent de celui du Tian Shan nord au Paléozoïque supérieur (Wang et al., 2006). Le bloc de Yili est séparé du STS par la faille de Narat et le CTS est séparé du STS par la faille de Baluntai (Fig. 1.2.7 ; (e.g. Charvet et al., 2011).

Le STS se différencie des autres blocs par une plus grande présence d'unités ophiolitiques, reliques de l'Océan du Tian Shan Sud (Dévonien) : Kumux et Hongliuhe dans le Tian Shan est (Shu et al., 2002 ; Laurent-Charvet, 2001 PhD ; Guo et al., 2002), et Heiyingshan et Aheqi à l'ouest (Wang, 2006 ; Wang et al., 2007 ; Wang et al., 2008). On y observe également des roches de type schistes bleus et des granulites de haute-pression d'âge Carbonifère inférieur (Gao et al., 1995 ; Wang et al., 1999) ; Shu et al., 2004 ; Liu & Hao, 2006 ; Shu et al., 2007). On y distingue deux zones : la partie sud est composée d'un socle de même type que celui de la marge nord du Tarim (socle protérozoïque gneissique), et la partie nord est principalement formée de mélanges ophiolitiques. La partie sud comprend des sédiments paléozoïques en partie métamorphisés (sédiments de plate-forme ordoviciens et antérieurs (Carroll et al., 1995), et est structurée en larges nappes à vergence nord. La partie nord, quant à elle, se compose principalement d'un mélange d'ophiolites à matrice schisteuse d'âge Silurien supérieur (Ma et al., 1993) et la matrice comporte des blocs de roches ultrabasiques et

basiques (diabases, roches siliceuses, radiolarites, marbres). Le tout est structuré en grandes écailles à vergence nord (Fig. 1.2.15, 1.2.16).

On notera également la présence de l'arc de Dananhu (ou Dananhu-Harlik) au nord-est du Tian Shan (Fig. 1.2.7), principalement composé d'un arc volcanique ordovicien à dévono-carbonifère et de turbidites, basaltes, cherts, etc. (Qin, 2000 ; Qin et al., 2002 ; Xiao et al., 2004b). Des roches calco-alkalines datant de l'Ordovicien-Silurien et pouvant montrer une origine de marge active, ont été observées le long de la marge sud du bassin de Turfan-Hami (Xiao et al., 2013). Ces roches permettront d'étudier l'évolution des sources des sédiments mésozoïques déposés dans le Bassin de Turfan, et par conséquent de reconstituer l'évolution de la chaîne.

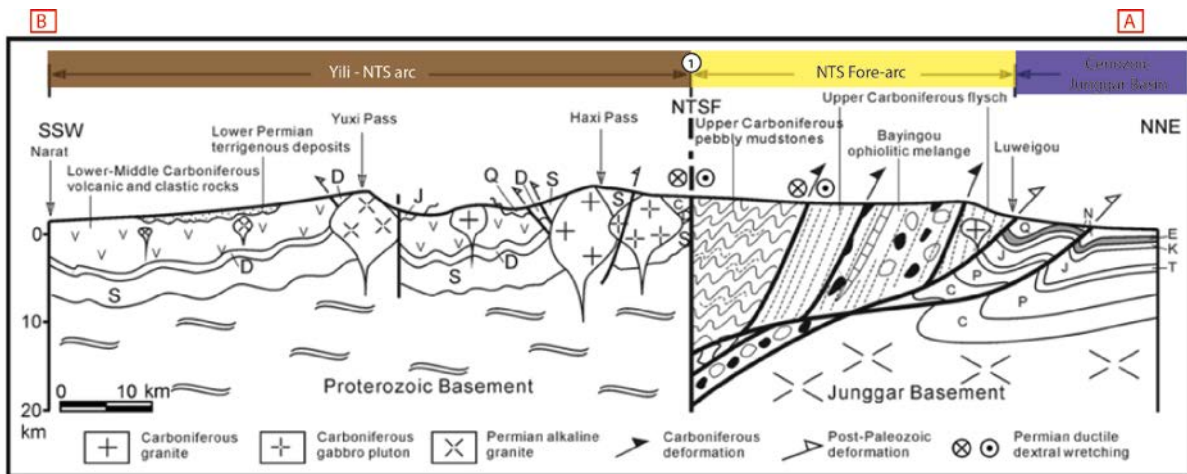


Figure 1.2.15 : Coupe interprétative (NNE-SSW) à travers le Tian Shan. Localisation : trait de coupe A-B sur la Figure 1.2.7. Légende, cf. Figure 1.2.7 – Le numéro 1 correspond à la Faille du Tian Shan Nord (NTSF), qui sépare le Bloc du Tian Shan Nord et le Bloc de Yili (Charvet et al., 2011, mis en couleurs).

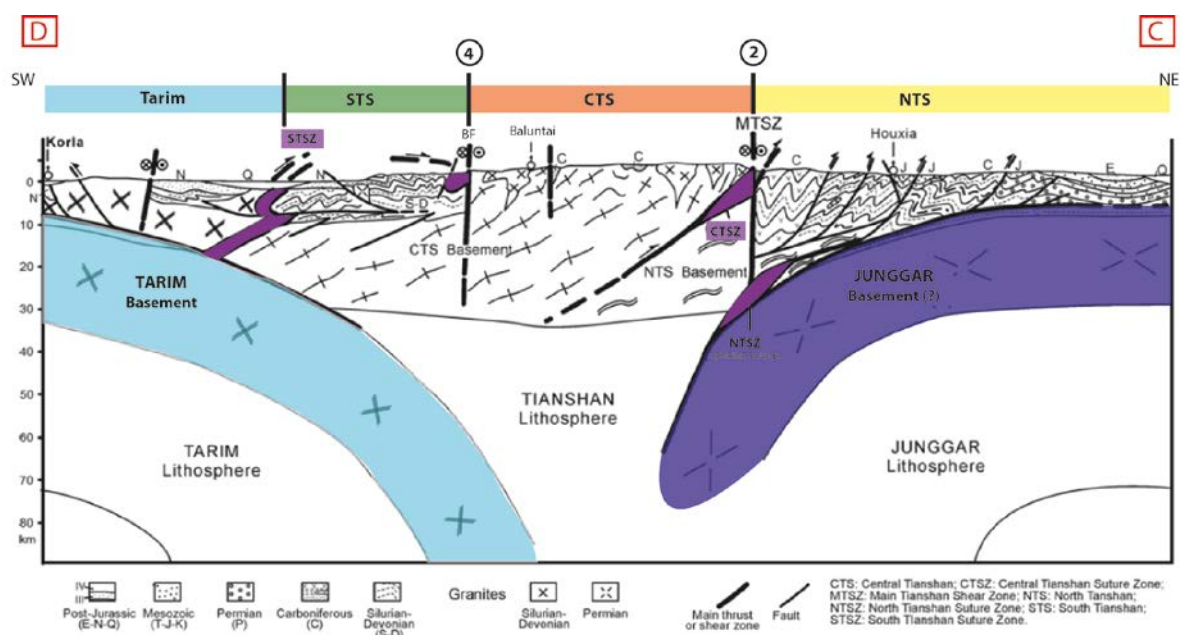


Figure 1.2.16 : Coupe interprétative (NE-SW) à travers le Tian Shan. Localisation : trait de coupe C-D sur la carte Figure 1.2.7. Légende, cf. Figure 1.2.7. Le numéro 2 correspond à la Zone de cisaillement principale du Tian Shan (MTSZ = Main Tian Shan Shear Zone) et le numéro 4 à la Faille de Baluntai (BF) (Charvet et al., 2011, mis en couleurs).

1.2.3 L'histoire paléozoïque supérieur : Bassins en transtension

En Asie Centrale, tous les blocs depuis le Tarim au sud jusqu'au bloc de Sibérie, sont accrétés au Permien inférieur à moyen (e.g. Charvet et al., 2007 ; Wang et al., 2007a). Dès la transition entre le Carbonifère et le Permien, des mouvements décrochant dextres ont lieu le long des failles majeures (faille de l'Irtysh dans l'Est-Junggar, le long de la MTSZ et de la faille de Narat) en raison des mouvements qui accommodent le cisaillement entre la Sibérie et le Tarim (Fig. 1.2.2) (e.g. Burtman, 1980 ; Levashova et al. 2003b ; van der Voo et al., 2006). Il en résulte la réactivation d'anciennes structures chevauchantes (e.g. Allen, et al., 1992, 1995 ; Chen, 1994 ; Cui, 1995 ; Dehandschutter, et al., 1997 ; Laurent-Charvet, 2001 ; Liu, et al., 1996 ; Melnikov, et al., 1997 ; Melnikov, et al., 1998 ; Shu, et al., 1999a ; Shu, et al., 1998).

Au Permien, les blocs formant la chaîne du Tian Shan sont amalgamés (Fig. 1.2.17) et les premiers sédiments se déposent dans les bassins bordant la chaîne avec des grès rouges et conglomérats discordant dans les bassins du Junggar et de Turfan (Wartes et al., 2002). Le dernier mouvement majeur de blocs a été enregistré par les données de paléomagnétisme et montre un mouvement de la plaque Kazakh (incluant l'ensemble Yili-Junggar ouest) avec un déplacement de

1 160 +/- 380 km vers l'Est par rapport aux blocs du Tarim et de Sibérie (considérés comme stables) (Wang et al., 2007b). Ces mouvements transtensifs entraînent la formation de failles décrochantes lithosphériques, telles que la zone de faille du Tian Shan nord (NTSF), la zone de cisaillement majeure du Tian Shan (MTSZ), la faille de Narat (NF) ou la ligne de Nikolaev (NL). Le long de ces grandes failles se forment des bassins permien de type « pull-apart », dont certains existent encore aujourd'hui (Yili, Bayanbulak, Turfan) (Fig. 1.2.2, 1.2.17). Ce contexte extensif est propice à la mise en place des granitoïdes permien à l'intérieur et le long des grandes zones de cisaillement, par exemple dans le Bloc de Yili (e.g. Wang et al., 2009), qui peuvent parfois être associés à des bassins (Allen & Vincent, 1997). A la même période, le bassin du Junggar est affecté par une phase d'extension générale (e.g. Allen et al., 1991b, 1995 ; Allen & Vincent, 1997 ; Qiu et al., 2005, 2008 ; Yang et al., 2013).

Ainsi au Permien supérieur, les mouvements décrochants permettent l'ouverture de bassins *pull-apart*, par exemple le long de la faille de Dalabute dans le NW du Junggar. Des épisodes volcaniques peuvent également être associés à ces mouvements décrochants (Fig. 1.2.17 ; Allen and Vincent, 1997).

Une phase de refroidissement et d'exhumation majeure (données AFT) est enregistrée du Permien au Jurassique moyen dans le socle du Tian Shan et correspond à une topographie positive significative (e.g. Jolivet et al., 2010). Les datations U/Pb sur zircons détritiques dans le Sud du Junggar ont également mis en évidence une large population (ca. 70 %) d'âge carbonifère et permien, montrant l'érosion de la chaîne tardi-paléozoïque depuis sa formation. Cette population est particulièrement présente dans les sédiments datant de la fin du Carbonifère jusqu'au Trias inférieur (e.g. Yang et al., 2013).

Au Permien inférieur, les derniers sédiments marins se déposent dans le bassin du Junggar, comme dans le bassin du Tarim (e.g. Carroll et al., 1995 ; Wartes et al., 2002 ; Bian et al., 2010). Dans le Junggar, ils sont recouverts par le produit de l'érosion des topographies actives (e.g. Tian Shan, Kelameili) et sont composés de dépôts terrigènes grossiers, qui progradent en direction du bassin (e.g. Carroll et al., 1990). Les dépôts permien sont ensuite dominés par des environnements lacustres couvrant une large partie du Junggar sud et du bassin de Turfan.

L'extension entre le bassin du Tarim et le Tian Shan sud a formé des sous-bassins relativement petits, mais profonds, tels que les bassins de Kuche et Yanqi (Fig. 1.2.17.b). Ils sont connectés avec l'océan Paléotéthys au Sud. Un peu plus au Sud (*Kalpin uplift*), les dépôts marins sont recouverts par des dépôts fluviaux du Permien inférieur, qui sont inter-digités avec des coulées basaltiques et indiquent également un contexte tectonique extensif (Carroll et al., 2001 ; Liu et al., 2014). La source de ce magmatisme est comparable à celle des autres coulées basaltiques datées du

Permien inférieur et présentes dans le reste du bassin du Tarim (cf. 1.1.1.1.2.1 LIP Tarim). Dans le sous-bassin de Kuqa, des rhyolites de plusieurs kilomètres d'épaisseur ont été datées à 282 ± 2 Ma (daté par Ar/Ar ; références dans Hendrix et al., 1992) et plus récemment la suite volcanique a été datée par ICP-MS à 295.0 ± 2.8 Ma (Liu et al., 2014). Les déplacements sur les failles décrochantes durent jusqu'au Permien supérieur - Trias inférieur (ca. 290-250 Ma, Fig. 1.2.17.b ; Laurent-Charvet et al., 2002 ; Shu et al., 2002 ; Wang et al., 2007a ; de Jong et al., 2009). Ils ont probablement cessé en raison de la fermeture de l'océan Paléotéthys et du régime compressif qui affecte la marge sud de l'Asie à cette époque (cf. 1.1.1.2 Asie : Trias) et particulièrement l'accrétion du bloc de Qiangtang (e.g. Allen et al., 1991b ; Wartes et al., 2002 ; De Grave et al., 2007 ; Roger et al., 2008, 2010, 2011).

Au Permien, le climat est probablement chaud et humide avec des variations saisonnières dans le bassin du Junggar (Thomas et al., 2011), alors que des conditions plus arides règnent dans le bassin du Tarim (Zhu et al., 2005b). La limite entre le Permien et le Trias correspond à une période d'aridification, comme l'attestent les nombreux paléosols observés dans le bassin de Turfan (e.g. Thomas et al., 2007, 2009). Cette période correspond également à un événement d'extinction de masse à l'échelle globale, que certains auteurs associent à un changement climatique majeur et qui reste toutefois très discuté (e.g. Rees, 2002 ; Retallack, 2013).

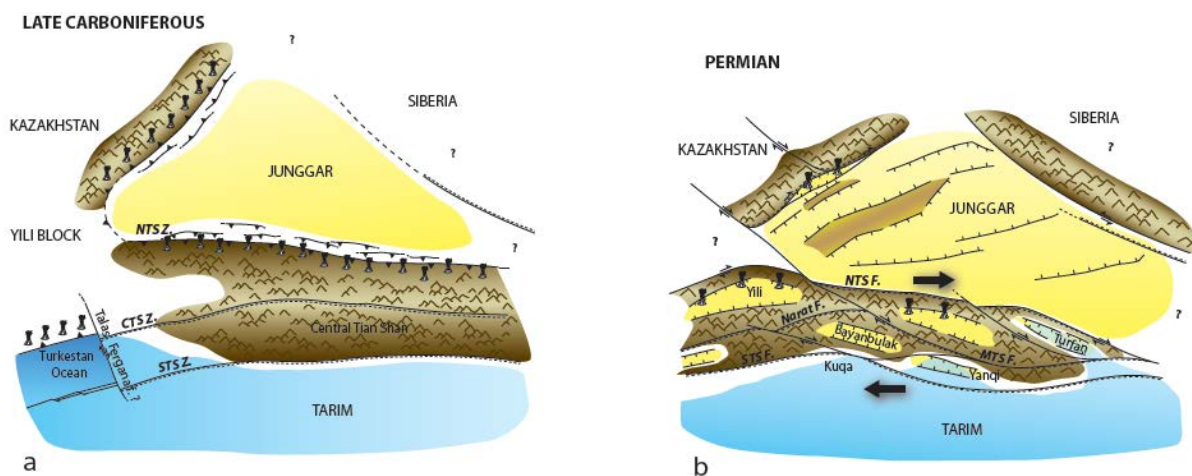


Figure 1.2.17 : Cartes paléogéographiques de la chaîne de Tian Shan et ses bordures nord et sud. a) Au Carbonifère supérieur, l'Océan du Tian Shan Nord est fermé et le régime général dans le Tian Shan est compressif. b) Au Permien, les mouvements en transpression permettent l'ouverture des bassins intramontagneux, tel que le Bassin de Bayanbulak (Jolivet et al., 2013c).

1.2.4 L'histoire mésozoïque : Erosion de la chaîne paléozoïque

Dès le début du Trias, la déformation due à l'orogénèse cimmérienne se propage dans la zone du Tian Shan et induit un contexte tectonique compressif. Les données de thermochronologie basse-température enregistrent une phase majeure de refroidissement, débutant au Trias inférieur et qui dure jusqu'au Jurassique moyen. Cette phase correspond à la réactivation et à l'érosion de la topographie permienne (e.g. Dumitru et al., 2001 ; De Grave et al., 2007 ; Jolivet et al., 2010). Les hauts de socle paléozoïques situés dans la partie centre-ouest du bassin du Junggar poursuivent leur exhumation pendant le Trias (Liu et al., 2006).

Dans les bassins du Junggar et de Turfan, les dépôts du Trias inférieur contiennent des dépôts fluviatiles rouges, et localement des dépôts conglomératiques de cônes alluviaux (e.g. Hendrix et al., 1992). Dans le sous-bassin de Kuqa, les dépôts sont dominés par des conglomérats de cônes alluviaux (Hendrix et al., 1992), aux provenances diverses, mais toujours issus de l'érosion du Tian Shan ancestral comme l'indiquent les paléo-courants (sud-est au sud-ouest), dans un contexte de compression générale et de surrection de la chaîne (e.g. Li et al., 2004 ; Liu et al., 2013b). Ces données montrent que la chaîne du Tian Shan constitue le relief à l'érosion à ce moment-là (e.g. Hendrix et al., 1992 ; Li et al., 2004 ; Liu et al., 2013b). Le climat de cette époque semble être plus aride que précédemment (Li et al., 2004). A partir du Trias moyen, les dépôts alluviaux et fluviaux sont plus courants dans les bassins du Junggar sud, Turfan et Tarim nord (Hendrix et al., 1992), ainsi que les dépôts de deltas lacustres et de plaines alluviales distales sur la coupe de Totoun dans le Junggar sud (Hornung & Hinderer, 2011 ; Sha et al., 2011), sur le bord nord du bassin du Junggar (Bian et al., 2010) et dans le Tarim nord (Li et al., 1995).

A partir du Trias moyen, la subsidence tectonique s'arrête, comme l'indiquent l'arrêt des discordances et la diminution des taux de sédimentation (Yang et al., 2013). Seul le bassin de Turfan reste dans un contexte extensif jusqu'au Trias moyen et la subsidence thermique y est forte (Shao et al., 1999). L'érosion de la topographie cimmérienne (qui se superpose probablement à la topographie permienne héritée) débute au Trias moyen et constitue le début d'une phase de pénéplanation majeure (Dumitru et al., 2001 ; Jolivet et al., 2010). Dans le bassin du Tarim, la composition des sables du Trias moyen au Jurassique moyen reflète une large érosion de la chaîne, accompagnée d'une pénéplanation. Le climat a été interprété comme étant chaud et humide, à partir des analyses de paléosols par exemple (e.g. Li et al., 2004 ; Ashraf et al., 2010), mais également grâce à l'analyse des dépôts de charbon et de morceaux de bois, qui indiquent la possible présence d'une forêt dans l'ouest du Junggar (McKnight et al., 1990). D'après ces mêmes interprétations, la fin du Trias et le début du Jurassique correspondent à un maximum d'humidité.

Ce contexte permet le dépôt de niveaux de charbon dans un environnement relativement stable tectoniquement. En effet, au Trias moyen, une large phase de formation de séquences à charbon débute dans les trois bassins, et dure jusqu'au Jurassique moyen (e.g. Hendrix et al., 1992). Dans les trois bassins (Junggar, Tarim et Turfan), les paléo-courants sont très dispersés et n'indiquent pas de direction unique (Xue et al., 2000).

Le Bassin de Turfan se déconnecte du Bassin du Junggar durant le Trias. Il est difficile de déterminer à quel moment exactement ce changement a lieu. En revanche, les mesures de paléo-courants dans le nord du bassin montrent qu'à partir du Jurassique inférieur les courants sont orientés vers le sud, indiquant le début de l'exhumation du Bogda Shan (Greene et al., 2001, 2005).

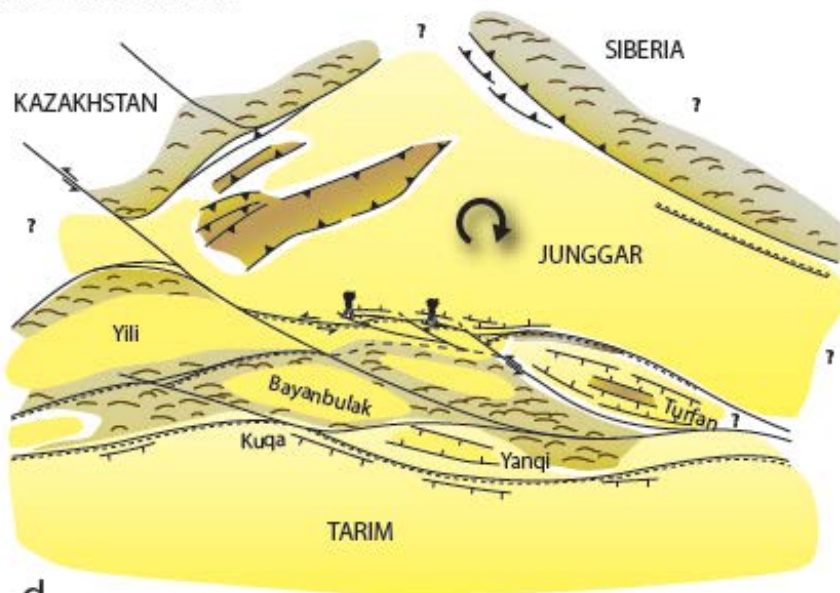
Le Jurassique est caractérisé par une diminution continue du relief du Tian Shan. En Asie Centrale, le régime tectonique de cette période est extensif (cf. 1.1.2) Asie - Mésozoïque). À l'ouest, le grand bassin du Caucase et le Bassin Caspien Sud s'ouvrent, respectivement au Jurassique inférieur et moyen (e.g. Zonenshain et al., 1990 ; Brunet et al., 2003). Au nord-est, une série de petits bassins jurassiques se forment, du sud-est de la Sibérie jusqu'au nord de la Chine (e.g. Zorin 1999 ; Donskaya et al., 2008 ; Daoudene et al., 2013 ; Jolivet et al., 2013b). Dans le bassin du Junggar, des signes d'extension (dans la partie sud du bassin) coexistent avec des signes de compression liée à la rotation horaire du bloc du Junggar (Fig. 1.2.18 ; Hou et al., 2009). Les bordures nord-ouest et nord-est du bassin sont sujettes à des mouvements chevauchants du Jurassique jusqu'au Crétacé inférieur (e.g. Allen et al., 1991b ; Liu et al., 2006). Les hauts de socle formés à la fin du Paléozoïque dans la partie centre-ouest du Junggar sont toujours actifs au Jurassique et Crétacé inférieur (e.g. He et al., 2008 ; Hou et al. 2009). Pourtant, de petits bassins transtensionnels se forment le long de la bordure sud du bassin du Junggar durant le Jurassique inférieur. Ils sont associés à une activité volcanique dont les dépôts de tuffs ont été observés (cf. Chapitre 2 ; e.g. Wang & Gao, 2012 ; Yang et al., 2013). Ces mêmes bassins ont enregistré des faunes d'eau saumâtre, probablement due à des incursions marines venant de l'océan Téthys nord, à l'ouest (Fig. 1.2.18 ; e.g. Sha et al., 2011). La sédimentation dans le bassin de Turfan semble être contrôlée par un régime extensif durant le Jurassique inférieur (e.g. Shao et al., 1999, 2003). Des failles normales ont également été observées sur la bordure nord du bassin du Tarim (e.g. Allen et al., 1991a).

EARLY JURASSIC



c

MIDDLE JURASSIC



d

Figure 1.2.18 : Cartes paléogéographiques de la chaîne de Tian Shan et ses bordures nord et sud. c) Au Jurassique inférieur, de possibles incursions marines proviennent de l'océan Téthys nord, à l'ouest – d) Au Jurassique moyen (Jolivet et al., 2013c).

A partir de la fin du Jurassique moyen, les données de thermochronologie basse température enregistrent une diminution marquée des taux de refroidissement du socle dans le Tian Shan. Elle marque le début d'une longue période de stabilité thermique, associée à une quasi-pénéplanation de la chaîne du Tian Shan (Dumitru et al., 2001 ; Jolivet et al., 2010). Ces surfaces continuent aujourd'hui d'être exhumées et sont observables quasiment partout en Asie Centrale, depuis le Tibet nord (Jolivet et al., 2001 ; Roger et al., 2011) jusqu'en Mongolie (Vassallo et al., 2007 ; Jolivet et al., 2007) et en Sibérie (De Grave et al., 2008 ; Jolivet et al., 2009, 2011, 2013a) (cf. 1.1.2) Mésozoïque moyen - supérieur).

A la même époque, l'accommodation dans les bassins sédimentaires est significative et plusieurs milliers de mètres de sédiments se déposent dans les bassins décrits précédemment (e.g. Hendrix et al., 1992). Particulièrement les sédiments jurassiques sont constitués de puissants dépôts. Dans la zone sud du bassin du Junggar, le principal centre de dépôt est quasiment confondu avec le point principal de subsidence, qui est situé près de Changji à proximité de la ville d'Urumqi (dans Yang et al., 2013, d'après Fang et al., 2005) (Fig. 1.2.19).

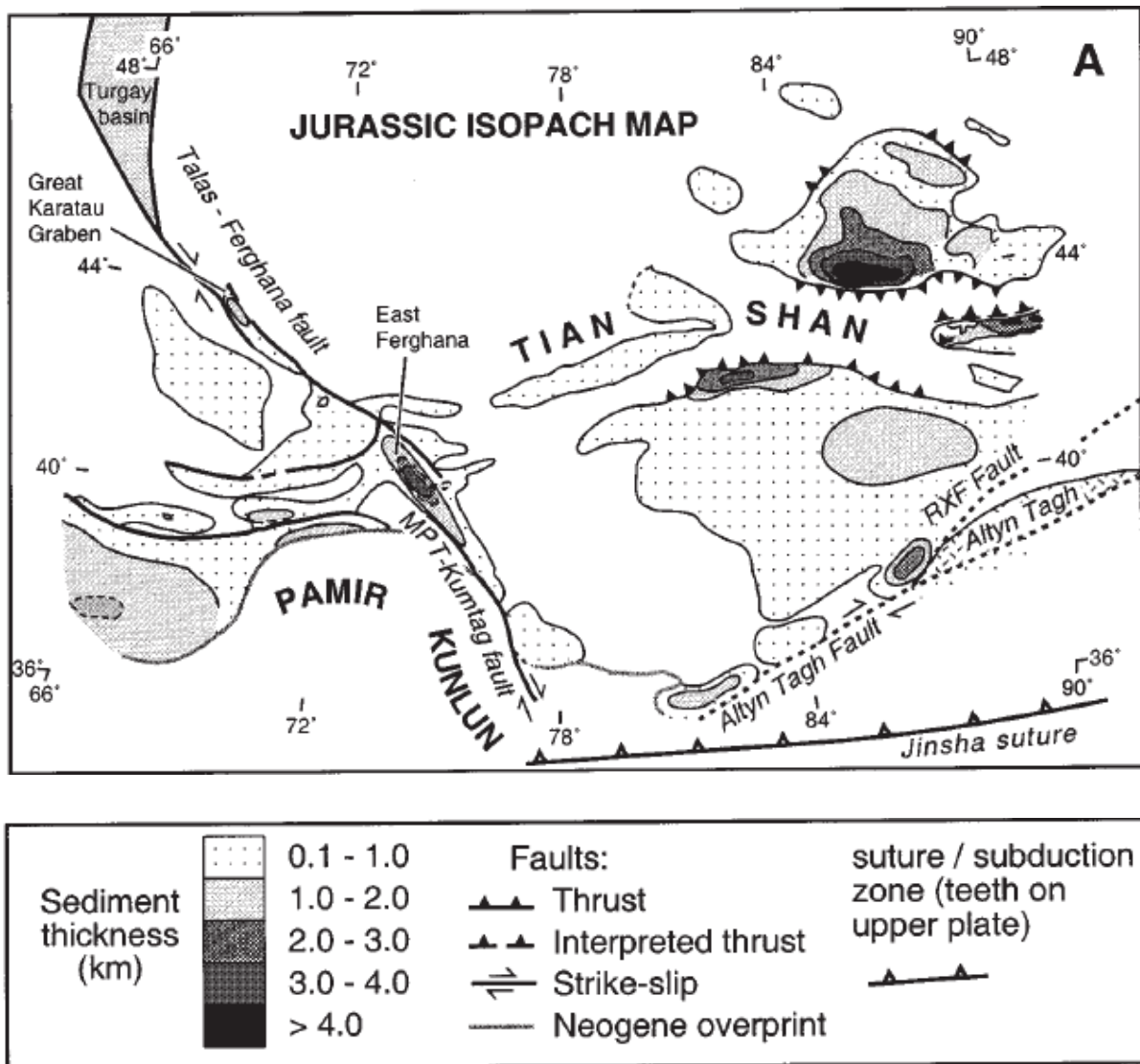


Figure 1.2.19 : Carte des isopaques représentant les dépôts jurassiques en Asie Centrale. Aucune restauration palinspatique n'a été réalisée. RXF = Faille de Ruoqiang-Xingxinxia ; MPT = Main Pamir thrust. (Sobel, 1999).

Tout au long du Jurassique, aucun changement significatif de direction n'a été observé dans les paléo-courants mesurés dans les dépôts tout autour de la chaîne. Dans le sous-bassin de Kuqa (Tarim nord), les mesures indiquent globalement une direction vers le sud, et sur la coupe de la Manas (Junggar sud), la principale direction observée est le nord. Ces directions montrent l'existence d'un haut topographique dans le Tian Shan durant le Mésozoïque (Hendrix et al., 1992). Par ailleurs, le massif du Bogda Shan, dont l'exhumation a débuté au Jurassique inférieur, est érodé et les paléo-courants de part et d'autre de la chaîne montrent des directions vers le bassin du Junggar au nord (Shao et al., 1999) et vers le bassin de Turfan au sud (e.g. Greene et al., 2001, 2005).

Dans les trois bassins, les dépôts du Jurassique inférieur (J1) se déposent d'abord dans des milieux fluviaux et sont accompagnés d'épais niveaux de charbon (Formation de Badaowan : environ 800 m d'épaisseur dans le Sud du bassin du Junggar et 400-500 m dans le bassin de Turfan), puis dans des environnements deltaïques sans niveaux épais de charbon (Sangonghe Formation : 10-480 m). Dans le bassin du Junggar, des incursions marines ont été supposées au Jurassique inférieur (Formation de Badaowan) sur la base des faunes et flores d'eau saumâtres trouvées sur la coupe de Haojiagou dans la Vallée de la Toutun (Fig. 1.2.18 ; e.g. Sha et al., 2011 ; Pan et al., 2013). L'origine de ces eaux saumâtres peut être discutée et pourrait être liée à un environnement plus aride, qui favoriserait l'évaporation et la concentration des eaux. L'abondance de sédiments riches en fer et en concrétions ferrugineuses indique de fortes précipitations et par conséquent un environnement chaud et humide, ponctué par des périodes plus sèches (Sha et al., 2010 ; Ashraf et al., 2010). Le Jurassique moyen (J2) consiste en des dépôts fluviaux (Formation de Xishanyao), qui contiennent les derniers et plus épais niveaux de charbon jurassique dans le bassin du Junggar (répartis sur plus de 1000 m de séries sédimentaires dans le Junggar sud et 100 – 500 m dans les autres bassins). La Formation de Toutunhe qui suit, est composée de 500 – 700 m de dépôts fluvio-lacustres. Le climat au Jurassique moyen est chaud et humide, mais comporte aussi des saisons sèches. A partir du Bathonien, les premiers signes de saisons sèches apparaissent à l'échelle du bassin du Junggar (Ashraf et al., 2010). Les dépôts du Jurassique supérieur (J3) se ressemblent dans les trois bassins (Junggar, Tarim et Turfan). Ils se composent des dépôts fluviaux rouges de la Formation Qigu (en général 300 – 500 m), recouverts des puissants dépôts alluviaux de la Formation de Kalazha (25 – 800 m). Seul dans le bassin de Turfan, la Formation de Kalazha qui est caractérisée par ses impressionnants dépôts conglomératiques dans les autres bassins, ne comporte que de rares séquences conglomératiques, dont l'épaisseur décroît du nord vers le sud (Shao et al., 1999). Elles y sont remplacées par de puissants dépôts gréseux (Sha et al., 2010). Une activité volcanique locale est soupçonnée, en raison des cinérites et grès tuffacés intercalés dans les dépôts sédimentaires de la partie inférieure de la Formation Qigu et dans le « Shishugou Group », observés respectivement dans le Junggar sud et dans le Nord-Est du Junggar (e.g. Sha et al., 2010 ; Yang et al., 2012 et les références associées). Le Jurassique supérieur marque un réel changement vers des conditions climatiques arides, qui peuvent être interprétés comme de très fortes précipitations ponctuelles, alternant avec des saisons particulièrement arides. Cette interprétation repose également sur des études histologiques effectuées sur des os de dinosaures (e.g. Hendrix et al., 1992 ; 2000 ; Ashraf et al., 2010 et nombreuses références associées). Cette hypothèse est également soutenue par les mégadunes éoliennes observées dans la Formation Shishugou supérieur (=J3, équivalent de la Formation de Kalazha) dans le nord-est du Junggar (Eberth et al., 2001). Vincent & Allen (2001) décrivent également des dépôts éoliens dans le nord-est du Junggar et Bourquin et al. (in review) des dépôts

éoliens sur la section de la Totoun, dans le sud du Junggar. Certains auteurs les associent au Crétacé inférieur, ce qui n'est pas surprenant puisqu'il est difficile de dater ces séries.

Dans le bassin du Junggar, les séries du Crétacé inférieur reposent généralement de manière discordante sur le Jurassique supérieur, traduisant une activité tectonique continue à cette période (Hendrix et al., 1992, 2000 ; Vincent & Allen, 2001 ; Sha et al., 2011 ; Bourquin et al. (in review). Les datations U/Pb sur zircons détritiques de la marge sud du bassin du Junggar montrent des évidences de cannibalisme (ou recyclage des sédiments précédemment déposés). Les bords du bassin ont migré vers le nord et la surface de celui-ci a considérablement diminué (e.g. Fang et al., 2006a ; Yang et al., 2013). Ceci correspond à une phase de déformation tectonique et de création de reliefs dans les piémonts nord du Tian Shan (Yang et al., 2013). Cette période correspond également au maximum de formation des hauts de socle au centre du bassin du Junggar (Hou et al., 2009).

Au Crétacé, la marge sud du bassin commence à être inversée, ce qui est généralement interprété comme une conséquence de la collision du bloc de Lhasa au Sud (cf. 1.1.3.2) Crétacé - Collision du bloc de Lhasa ; e.g. Hendrix et al., 1992 ; Gu, 1996 ; Fang et al., 2006b). Pourtant, les données de thermochronologie n'enregistrent qu'une longue période sans phase d'exhumation majeure. Il s'agit plutôt d'un réajustement mineur des structures, sans création de topographie majeure, où le bassin conserve ses caractéristiques. En résumé, depuis le Trias moyen-supérieur jusqu'au Paléogène, la structure en « *basin and range* » a dominé la partie sud du bassin du Junggar. De la même manière, un soulèvement est mis en évidence sur la marge nord du bassin du Tarim grâce aux datations U/Pb sur les zircons détritiques. Ceci entraîne l'augmentation du relief entre le bassin et la chaîne (Li & Peng, 2010).

De manière générale, les déformations affectant les piémonts du Tian Shan au Crétacé inférieur sont interprétées comme liées à la collision du bloc de Lhasa au Sud (e.g. Hendrix et al., 1992 ; Gu, 1996 ; De Grave et al., 2007 ; cf. 1.1.3.2) Crétacé - Collision du bloc de Lhasa). Pourtant, des régions telles que l'Est du Tibet ou la marge sud du bassin du Tarim, qui sont plus proches de la zone de collision n'enregistrent que très peu de déformation due à cette collision, ce qui remet en cause cette interprétation (e.g. Coward et al., 1988 ; Jolivet et al., 2001, 2010 ; Roger et al., 2008, 2010, 2011).

D'autre part, les données de thermochronologie basse température obtenues sur le socle de la chaîne du Tian Shan indiquent que la phase quasi-isotherme débutée au Jurassique moyen, a duré jusqu'à la fin du Crétacé, ce qui contredit la présence d'une phase d'exhumation au Crétacé inférieur

(Dumitru et al., 2001 ; Jolivet et al., 2010). En revanche, plus vers l'Ouest (au Kirghizstan) la phase isotherme ne se produit qu'à partir de la fin du Crétacé et suit une période d'exhumation limitée (e.g. De Grave et al., 2007 ; Glorie et al., 2010).

Les dépôts du Crétacé inférieur correspondent à des dépôts fluvio-lacustres dans le sud du bassin du Junggar (cf. faune d'eau douce, cf. références citées dans Hendrix et al., 1992) et à des dépôts fluviaux dans le sous-bassin de Kuqa (Tarim nord) (Hendrix et al., 1992). Dans le bassin de Turfan, Shao et al. (1999) décrit également des environnements fluvio-lacustres, alors que Hendrix et al. (1992) interprètent des systèmes de rivières en tresses.

Au Crétacé supérieur, les environnements lacustres laissent place à des deltas, des systèmes de rivières en tresse, voire des cônes alluviaux (Hendrix et al., 1992 ; Shao et al., 1999 ; Wang et al., 2013) dans les trois bassins (Junggar, Turfan et Tarim).

La transition entre la fin du Crétacé et le début du Paléogène est marquée par un épisode de réactivation majeure des structures paléozoïques telle que la ligne de Nikolaev (Fig. 1.2.20). Le long de la faille de Narat, les données de thermochronologie basse température (AFT et (U-Th)/He) indiquent une phase de refroidissement rapide vers 65-60 Ma. Ces âges sont également observés dans la chaîne du Bogda Shan et indique son exhumation (Wang et al., 2008 ; Jolivet et al., 2010). Ces réactivations locales pourraient être liées aux effets de grandes longueurs d'ondes de l'accrétion des ensembles structuraux au sud-ouest (Arc du Kohistan-Dras, bloc de Kaboul par exemple) ou à la fermeture finale de l'océan du Mongol-Okhotsk au nord-est (cf. 1.1.3.3 Transition Crétacé – Tertiaire ; e.g. Searle, 1991 ; Hendrix et al., 1992 ; Wang et al., 2008 ; Jolivet et al., 2010). Le relief est non-négligeable à ce moment (Jolivet et al., 2013c).

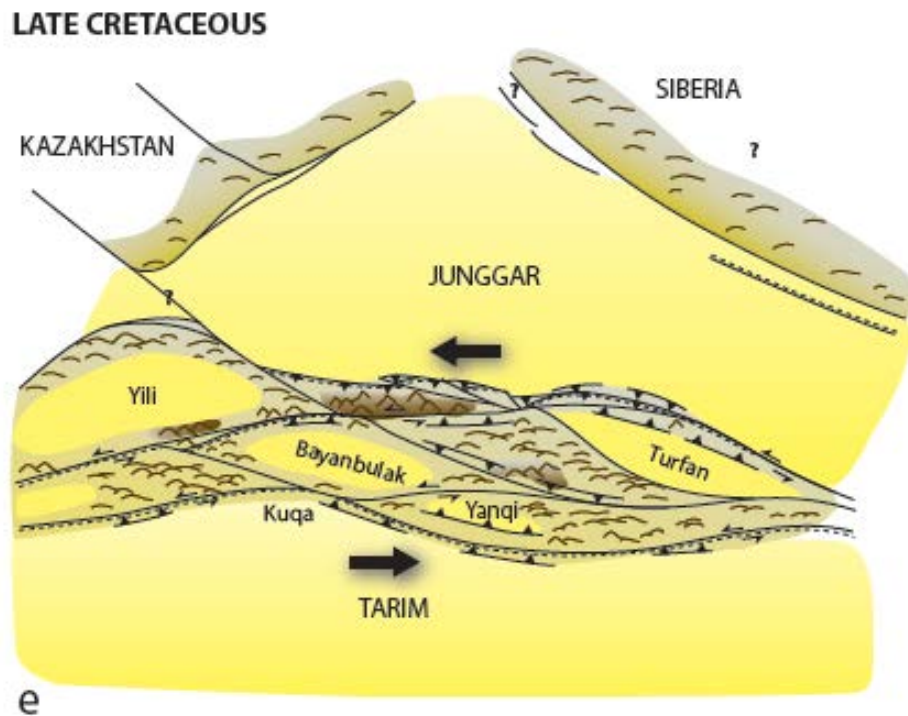


Figure 1.2.20 : Cartes paléogéographiques de la chaîne de Tian Shan et ses bordures nord et sud au Crétacé supérieur. La chaîne est réactivée et le régime général est compressif (Jolivet et al., 2013c).

L'observation de paléosols permet d'affirmer que le climat était aride à partir de la fin du Jurassique et au Crétacé dans l'ensemble de l'Asie Centrale (e.g. Eberth et al., 2001; Li et al., 2004; Jiang et al., 2008).

Le début du Paléocène semble être marqué par une période de calme tectonique, du Kirghizstan à l'ouest (e.g. Fortuna, 1983 ; Cobbold et al., 1994 ; Glorie et al., 2010 ; Macaulay, 2013) jusque dans le Tian Shan chinois à l'est (e.g. Jolivet et al., 2010). Cette période précède la réactivation de la chaîne qui résulte de la collision Inde-Asie et qui débute au Miocène (e.g. Dumitru et al., 2001 ; Bullen et al., 2003 ; Sobel et al., 2006a, 2006b ; Wang et al., 2009 ; De Grave et al., 2012, 2013 ; Macaulay et al., 2013 ; Macaulay, 2013).

Les environnements de dépôt varient depuis les deltas jusqu'à la plaine alluviale. Le climat est supposé être aride à semi-aride, influencé par une forte saisonnalité, comme interprété durant le Miocène moyen (e.g. Sun and Wang, 2005; Bougeois et al., 2014).

A partir de la fin de l'Eocène et de l'Oligocène, les premières déformations compressives à grandes échelles ont lieu et sont contemporaines de la formation du Pamir (au sud-ouest du Tian Shan) (e.g. Burtman, 1993 ; Arrowsmith & Strecker, 1999 ; Burtman, 2000).

**Sedimentological analysis and paleogeographic reconstitutions of the Mesozoic
Chinese Tian Shan area**

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INTRODUCTION

The Tian Shan (Central Asia) is a Cenozoic intra-continental range that develops in an area submitted to recurrent tectonic deformation since Late Paleozoic. The main lithospheric structures of the Tian Shan area were set during the Paleozoic final accretion of the Central Asian Orogenic Belt (CAOB) (e.g. Charvet et al., 2011; Windley et al., 2007; Wang et al., 2009; Alexeiev et al., 2011, Glorie et al., 2011a). The Tertiary evolution of the range is driven by the far field effects of the India-Asia collision and in that respect, has been largely described by tectonic, sedimentological, geochronological and geomorphic studies (e.g. Tapponnier & Molnar, 1979; Allen et al., 1994; Avouac et al., 1993a, Charreau et al., 2009b; Jolivet et al., 2010; Glorie et al., 2010; De Grave et al., 2013).The

Mesozoic evolution of the Tian Shan area, between the end of the Late Paleozoic orogeny and the onset of the Tertiary reactivation, remains however poorly understood and controversial. Low temperature thermochronology and tectonic data within the basement of the Chinese Tian Shan suggest a near to complete absence of exhumation and tectonic movement from Middle Jurassic to Early Cenozoic (Dumitru et al., 2001; Jolivet et al., 2010). However, detrital geochronology within both the Tarim and Junggar basins indicates sediment recycling during the same period, suggesting vertical movements and erosion (e.g. Li and Peng, 2010; Yang et al., 2013; Liu et al., 2013). Similarly sedimentological analyses in the Tarim and Junggar foothills of the range also indicate some recurrent tectonic events (e.g. Allen & Vincent, 1997 ; Hendrix et al., 1992, 2000; Sobel, 1999). In order to solve this apparent paradox, we aimed to reconstruct the evolution of the Mesozoic topography in the Tian Shan area.

The reconstruction of the paleogeography and associated topography can be developed from two different points of view (e.g. Reiners, 2007): either in looking at the material that has not been eroded (the basement) or in looking at the eroded material that was deposited further away (in the basins). The first approach consists in studying the thermal history recorded in the range, from which the main exhumation phases of the basement can be inferred. The second approach is based on the analysis of the sedimentary deposits accumulated in surrounding basins which record the evolution of the paleo-topography growth and its erosion.

The bedrock-based approach rests upon geochronological methods such as fission track and (U-Th)/He analysis. It provides exhumation rates of the basement and statistical models describing the thermal evolution of the range. This allows inferring the age of the main tectonic phases affecting an area and their relative intensity. The vertical resolution of the inferred cooling histories remains limited (e.g. Ehlers & Farley, 2003) and makes it difficult to infer a detailed topography. Furthermore the information contained in the minerals can be erased either by erosion and loss of the rock, or through burial above the closing temperature.

On the other hand, the products of the erosion of a relief can be collected in the adjacent sedimentary basins and constitute an available and long-lasting data source. The first step when studying these deposits is to determine sedimentary facies, in order to infer the depositional environments and to observe their possible lateral variations. The volume of sediments reflects the quantity of erosion (depending on climatic conditions and the lithology of the eroded rocks) and the depositional environments show the distance to the relief. This is usually accompanied by the analyses of paleo-current directions, which indicate the local slope direction in the basins. Several sections spread over the study area are required in order to draw paleogeographic maps. For that

reason, the sedimentological approach requires studying numerous sections in a sedimentary basin to increase the resolution of the inferred surrounding reliefs. Yet, the comparison and correlation of several sections can be challenging, especially if the sediment series are difficult to date. Indeed, the scarcity of biostrata and other markers (tuff, etc.) or the impossibility of dating the series with paleomagnetism can be problematic, as in the Tian Shan area.

In an endoreic basin for instance, the record of the tectonic evolution of an area can be biased e.g. by climatic changes, which can modify the base level of the sedimentary basin, and hence its record of the tectonic signal.

The geochronological analyses of detrital grains using U/Pb or $^{40}\text{Ar}/^{39}\text{Ar}$ depends on the content of zircons, micas and other minerals in the studied sediments. These minerals are relatively common, but can be absent from certain type of rocks (e.g., limestone, etc.). The called methods deliver a precise age of each grain, and enable us to infer the relative sediment sources and their evolution at a large scale, as long as the potential source rock areas are known and dated with geochronological methods, and the sediment deposition age is known. The use of low-temperature methods can be limited, because their temperatures of reset are easier to reach.

Thanks to the latter approach which rests upon the erosion products, it is possible to get information about completely eroded ancient reliefs. Yet it is very difficult to reconstruct the evolution of a paleo-topography only from this type of analyses. From the erosion zone to the deposition area, the transport of the eroded source rock is not linear. Sediments can be trapped in intermediate basins and re-eroded later in a further evolution stage of the reliefs (sediment recycling, e.g. Jolivet et al., 2013c). Moreover, some of the used minerals (such as zircons) are quite resistant to mechanical abrasion occurring during erosion and transport, and can be recycled several times. This will bias the interpretation and shows the necessity of combining this approach to other types of data.

These methods provide relatively accurate and reliable data sets, but need to be combined to each other in order to derive paleogeographic models. We propose to associate the bedrock and the basin-approaches with the aim of completing the information given by one approach, with the other. In that way, we should be able to discuss the vertical movements that have affected the study area through the Mesozoic and decipher the topographic evolution of the Tian Shan range.

In this study we present a detailed sedimentological analysis on six Mesozoic sections in both the foothills of the Tian Shan range and intra-mountain basins. The results are summarized on paleoenvironments reconstruction maps imaging the connection or isolation of the basins through

time, and hence on the location of reliefs. This analysis provides first-order insights on the topographic evolution of the area from Late Triassic to Early Paleogene. The results are combined to published low-temperature thermochronology data on the basement that provide the large-scale exhumation history of the range. The ultimate aim is to describe the Mesozoic tectonic and topographic evolution of the Chinese Tian Shan area and discuss this evolution in the framework of the main geodynamic processes that affected the Asian continent during that period.

GEOLOGICAL SETTING

1) TECTONICS

The Tian Shan range is one of the main intra-continental mountain ranges in Central Asia (Fig. 2.1a). Nowadays, it is surrounded by two major endoreic foreland basins: the Tarim Basin to the south and the Junggar Basin to the north. The range can be divided in three main regions: the northern, central and southern Tian Shan (Fig. 2.1.c). These regions correspond to different blocks which final accretion took place during Late Carboniferous – Early Permian times (e.g. Allen et al., 1993b; Gao et al., 1998; Wang et al., 2006; Charvet et al., 2007, 2011). The South Tian Shan block (STS), including some of the north Tarim margin is composed of Silurian gneiss and ophiolites (Zhou et al., 2001; Laurent-Charvet, 2001; Charvet et al., 2011). The Central Tian Shan – Yili block (CTS) is formed by an Ordovician volcanic arc associated to metamorphic Silurian flysh series lying on Proterozoic crystalline basement (e.g. Gao et al., 1998; Zhou et al., 2001; Dong et al., 2011). Finally the North Tian Shan terrane (NTS) consists in a series of Devonian to Early Permian island arcs accreted between the Tarim–Central Tian Shan and Junggar blocks during the Late Carboniferous–Early Permian closure of the Central Asian Ocean (e.g. Allen et al., 1993b; Charvet et al., 2011). The calc-alkaline arc series are intruded by Permian post-collision granitoids (Zhou et al., 2001; Liu et al., 2014)(Fig. 2.1b).

Permian

During the Permian, the Tian Shan area was affected by a major transtension episode. Palaeomagnetic data suggest that the Kazakh plate, including the Yili – West Junggar collage moved eastward by $1,160 \pm 380$ km respectively to stable Tarim and Siberia (Wang et al., 2007b). This motion generated lithospheric-scale strike-slip fault zones such as the North Tian Shan Fault, the

Main Tian Shan Shear Zone, the Narat Fault or the Nikolaev Line and partially controlled the emplacement of the Tian Shan Permian granitoids (Wang et al., 2009) (Fig. 2.1b). Strike-slip motion lasted from Early Permian to Early Triassic (c.a. 290 – 250 Ma; Laurent-Charvet et al., 2002; Shu et al., 2002; Wang et al., 2004, 2007a, de Jong et al., 2009). Low temperature thermochronology within the Tian Shan basement indicates a strong cooling – exhumation phase implying a major positive topography also suggested by detrital zircon U-Pb dating within the south Junggar and the North Tarim basins (Jolivet et al., 2010; Liu et al., 2013; Yang et al., 2013).

Pull-apart basins formed along those strike-slip fault zones, sometimes associated to volcanism (Allen and Vincent, 1997). Within the Tian Shan, the Yili, Bayanbulak and Turfan basins developed between the major strike-slip faults (Fig. 2.1b). At the same time, a general extension phase affected the Junggar Basin (e.g. Allen et al., 1991b, 1995; Allen and Vincent, 1997; Qiu et al., 2005, 2008; Yang et al., 2013). Evidence for normal faulting during Early Permian has also been reported by Carroll et al., 1999 from the Turfan Basin. While the Junggar basin became fully continental during Late Carboniferous – Early Permian, extension between the Tarim and South Tian Shan allowed the formation of deep, small-scale basins, such as the Kuqa and Yanqi basins (Fig. 2.1b). These connected with the Palaeo-Tethys Ocean to the south. The closure of the Palaeo-Tethys Ocean and the following Cimmerian collisions during the Permian – Early Triassic, and especially the accretion of the Qiangtang block, might have ended the transtension phase (e.g. Allen et al., 1991b; Wartes et al., 2002; De Grave et al., 2007; Roger et al., 2008, 2010, 2011).

Triassic

From Early Triassic, the far-field effects of the Cimmerian orogeny induced a compressive tectonic regime within the Tian Shan area. Low temperature thermochronology data indicate a significant cooling phase from Early Triassic to Middle Jurassic times corresponding to reactivation and erosion of the Permian topography (e.g. Dumitru et al., 2001; De grave et al., 2007; Jolivet et al., 2010). Late Paleozoic basement highs in the Junggar basin continued to uplift during the Triassic (Liu et al., 2006) and tectonic subsidence ceased by Middle Triassic (Yang et al., 2013). However, the Turfan basin remained extensional until the Middle Triassic (Shao et al., 1999). In the Tarim Basin U-Pb ages obtained on detrital zircons show similar age groups in the Lower to Middle Triassic deposits than in the Upper Permian ones. This can be interpreted as continuous erosion of the major positive topography formed during the Lower Permian and separating the south Junggar from the north Tarim basins (e.g. Liu et al., 2013). From Middle Triassic to Middle Jurassic, erosion of the Cimmerian topography marks the onset of a general planation phase, as suggested by low-temperature data (Dumitru et al., 2001; Jolivet et al. 2010) and sandstone petrography (e.g. Li et al., 2004).

Jurassic

The Jurassic period is characterized by a continuous decrease of the Tian Shan topography by erosion (e.g. Dumitru et al., 2001; Jolivet et al., 2010). During this period, Central Asia is affected by a general extensive tectonic setting. To the west, the Great Caucasus Basin and the South Caspian Basin opened respectively in Early and Middle Jurassic (e.g. Zonenshain et al., 1990; Brunet et al., 2003). To the northeast, a series of small-scale Late Jurassic basins are spread from southeast Siberia to North China (e.g. Zorin, 1999; Donskaya et al., 2008; Daoudene et al., 2013; Jolivet et al., 2013b). In the Junggar Basin, Jurassic extension and compression features co-existed due to the clockwise rotation of the Junggar Block (Hou et al., 2009). The northern and western edges of the basin show evidences of Jurassic to Early Cretaceous limited thrusting and the paleo-highs that initiated during the Late Paleozoic within the basin were still present during the Jurassic and Lower Cretaceous. However, along the southern edge of the Junggar Basin, small transtensive basins developed during the Lower Jurassic, associated to volcanic activity and brackish-water sedimentation (Sha et al., 2011; Wang and Gao, 2012; Yang et al., 2013). Similarly, the Turfan Basin was also controlled by extension during the Lower Jurassic (e.g. Shao et al., 1999; 2003; Allen et al., 1993a) and evidences of normal faulting are reported from the North Tarim margin (e.g. Allen et al., 1991a; Kuang et al., 2010).

Low temperature thermochronology data within the Tian Shan indicate a strong decrease in cooling rates from the late Middle Jurassic initiating a long period of thermal stability associated to a near planation of the previous reliefs (Dumitru et al., 2001; Jolivet et al., 2010). This planation is observed throughout most of Central Asia, from North Tibet (Jolivet et al., 2001; Roger et al., 2011) to Mongolia (Vassallo et al., 2007; Jolivet et al., 2007) and Siberia (De Grave & Van den haute, 2002; Buslov et al., 2008; Jolivet et al., 2009, 2011, 2013a).

Cretaceous

Within the Junggar Basin, the Lower Cretaceous series are mostly unconformably lying on the Upper Jurassic attesting of continuing tectonic activity during that period (Hendrix et al., 1992, 2000; Vincent and Allen, 2001; Sha et al., 2011). U/Pb dating on detrital zircons along the southern edge of the Junggar Basin shows evidences of sediment recycling, again consistent with tectonic deformation and relief building in the North Tian Shan foothills (Yang et al., 2013). The paleo-highs within the basin reached their climax during the Lower Cretaceous (Hou et al., 2009). Along the northern margin of the Tarim basin, a similar uplift is attested again by U/Pb dating on detrital zircons, increasing the differential topography between the basin and the relief (Li and Peng, 2010).

Those Early Cretaceous deformations within the Tian Shan foothills have often been associated with the collision of the Lhasa block to the south (Hendrix et al., 1992; Gu, 1996; De Grave et al., 2007). However, the apparently very small amount of deformation induced by the collision in eastern Tibet or around the southern margin of the Tarim Basin still questions this interpretation (e.g. Coward et al., 1988; Roger et al., 2008, 2010, 2011; Jolivet et al., 2010). Low temperature thermochronology data within the Chinese Tian Shan indicate that the near isothermal stage that initiated during the Middle Jurassic lasted until the Late Cretaceous contradicting the Early Cretaceous uplift phase (Dumitru et al., 2001; Jolivet et al., 2010). However, to the west in Kyrgyzstan, this isothermal phase occurs only from the Late Cretaceous, following a period of limited cooling (e.g. De Grave et al., 2007; Glorie et al., 2010; De Grave et al., 2013).

The Late Cretaceous – Early Paleogene transition is marked by a strong, localized reactivation of the Late Paleozoic inherited structures such as the Nikolaev Line. Apatite fission track analysis and (U-Th)/He data indicate a strong cooling event along the Narat Fault around 65 to 60 Ma as well as the onset of uplift of the Bogda Shan around 65 Ma (Wang et al., 2008; Jolivet et al., 2010) (Fig. 2.1b). This reactivation event is still poorly documented but might be linked to far-field effects of terrane accretions to the southwest (Kohistan-Dras Arc, Kabul Block) or to the final closure of the Mongol-Okhotsk Ocean to the northeast (e.g. Searle, 1991; Hendrix et al., 1992; Wang et al., 2008; Jolivet et al., 2010).

2) SEDIMENTOLOGY

Stratigraphy and biostratigraphy

The following descriptions are based on data from the literature. They mostly rely on the work of Hendrix et al. (1992) and have been completed by more recent studies such as Eberth et al. (2001), Shao et al. (1999) and Jiang et al. (2008). The formation boundaries rely on palynostratigraphical and biostratigraphical data (e.g. Eberth et al., 2001; Jiang et al., 2008; Ashraf et al., 2010; Bian et al., 2010; Sha et al., 2010; Hornung & Hinderer, 2011; Sha et al., 2011). However some stratigraphic boundaries still remain questionable (Table 1). In deed the age of some deposits can be determined if pollen and/or fossils have been well preserved. Pollen enabled the determination of series from Late Triassic (Huangshanjie Formation) to early Middle Jurassic (Xishanyiao Formation). Ashraf et al. (2010) did not find any pollen in Middle to Upper Jurassic strata (Xishanyiao to Kalazha Formations). Nevertheless, vertebrates have been described in other parts of

the Junggar Basin in the Totoun and Qigu Formations (e.g. Eberth et al., 2001; Ashraf et al., 2010 and references therein). Despite these indicators, the age of most of the Mesozoic formations remain difficult to constrain precisely.

The descriptions below summarize sediment deposition from the Triassic to the Paleogene in several basins adjacent to the Tian Shan range, mostly the southern margin of the Junggar Basin (S-Junggar), the Turfan and the Kuqa sub-basin in the northern margin of the Tarim Basin (N-Tarim) (Fig. 2.1; Table 2).

Triassic

The Lower Triassic series are dominated by braided fluvial redbed deposits in the S-Junggar and Turfan basins, locally with conglomeratic alluvial fan deposits (Hendrix et al., 1992; Greene et al., 2001; Vincent and Allen, 2001; Shao et al., 2003). Bian et al. (2010) also describe alluvial fans and deltaic deposits on the other Junggar Basin margins, associated to large lakes in the basin center. South of the range, in the Kuqa sub-basin, sediments are dominated by conglomeratic alluvial fan deposits (e.g. Hendrix et al., 1992). In this basin and its surroundings, Early Triassic paleocurrent directions show a southeast to southwest dispersal system, which shows a slope direction from north to south indicating a high to the north, i.e. the ancient Tian Shan range (Li et al., 2004).

During the Middle Triassic, alluvial and braided fluvial deposits are common in the S-Junggar, Turfan and N-Tarim Basins (Hendrix et al., 1992; Greene et al., 2001). Lacustrine delta and distal alluvial plain environments have also been observed on the northern edge of the Junggar Basin (Bian et al., 2010), locally on the southern edge of the Junggar Basin (Totoun Section, Haojiagou valley in Hornung & Hinderer, 2011, Sha et al., 2011), and in the N-Tarim Basin (Li et al., 1995). In each basin (S-Junggar, Turfan, N-Tarim) paleocurrent directions are scattered in all directions (Xue et al., 2000).

The Uppermost Triassic strata are comparable in the S-Junggar and N-Tarim Basins. They consist in alluvial conglomerates that shift into deltaic or lacustrine environments (e.g. Hendrix et al., 1992; Kuang et al., 2010). The Turfan Basin records braided-fluvial sand- and siltstones and associated floodplains deposits (Greene et al., 2001). The Upper Triassic period marks the onset of a widespread coal-bearing sediment deposition phase lasting up to the Middle Jurassic throughout the three basins (Hendrix et al., 1992; Jiang et al., 2008; Bian et al., 2010) (Table 2).

It is unclear when exactly during the Triassic the Turfan basin disconnected from the Junggar basin. However paleocurrent measurements in Early Jurassic series around the Bogda Shan, as well as provenance studies, show north-oriented transport directions in the northern foothills and south-

oriented directions in the southern foothills of the growing Bogda Shan. This indicates the occurrence of two separate basins in Early Jurassic (e.g. Hendrix et al., 2000; Greene et al., 2001; Shao et al., 2003; Greene et al., 2005).

Jurassic

The Jurassic deposits overlay unconformably the Triassic series throughout the three basins (Hendrix et al., 1992; Shao et al., 1999) (Table 2). In the S-Junggar Basin the main Lower Jurassic depositional environment evolves from a braided fluvial or alluvial plain system with thick coal beds (e.g. Badaowan Formation) to delta or shallow lakes with thinner and fewer coal beds (e.g. Sangonghe Formation, see Table 1). Locally, Early Jurassic deposits (e.g. Badaowan Formation) show some brackish-water fauna and flora on the Haojiagou section (about 50km southwest of Urumqi), on the Manas section (about 70km south of Shihezi), on the Anjihai section (about 80 km west of Shihezi) in the S-Junggar, and on the Tuziakeneigou section (about 5km west of Karamay) in the north-western part of the basin (Fig. 2.1). They are interpreted as evidences for local transgression events of the North Tethys Ocean from the west (Chen and Zhang, 2000; Sha et al., 2011; Pan et al., 2013). Fluvial delta and lacustrine deposits are reported from the Turfan Basin (Hendrix et al., 1992; Shao et al., 1999; Shao et al., 2003) and meandering fluvial series from the N-Tarim Basin (Hendrix et al., 1992; Jiang et al., 2008).

The Middle Jurassic is mostly composed of alluvial plain and deltaic or lacustrine deposits (Xishanyao Formation), which contain the topmost and thickest Jurassic coal-bearing strata of the S-Junggar (in alternation with silt- and sandstones in up to 1000m of deposits). These deposits are overlaid by fluvial lacustrine sediments (Totounhe Formation) (e.g. Hendrix et al., 1992; Eberth et al., 2001; Gao et al., 2003; Bian et al., 2010). The Turfan and N-Tarim Basin are characterized by swamp and meandering fluvial environments (Hendrix et al., 1992; Shao et al., 1999; Shao et al., 2003; Jiang et al., 2008).

The Upper Jurassic deposits are similar in the S-Junggar, Turpan and N-Tarim basins. Generally they are divided in redbed braided fluvial or alluvial plain deposits at the base (Qigu Formation) and thick alluvial fan conglomerate deposits at the top (Kalazha Formation). In the north-eastern part of the Junggar Basin, fluvial lacustrine and eolian deposits have been identified in the Kalazha Formation. The Kalazha Formation is generally considered to belong to the Upper Jurassic (e.g. Hendrix et al., 1992; Shao et al., 1999), but some authors assess this outstanding formation to

be of Lower Cretaceous age (e.g. Eberth et al., 2001; Vincent and Allen, 2001; Sha et al., 2010; Wang et al., 2013). The lack of reliable age constraints prevents solving this problem.

Local pink or white tuffaceous sandstones interbedded in the sediments are described from the Xidagou section and Santanghu Basin (north-east of the Turfan Basin) in upper Middle Jurassic series and from the Manas section in lower Upper Jurassic series (Sha et al., 2010). They are likely linked to volcanic activity in the surrounding region (Yang et al., 2013).

Throughout the Jurassic period, no significant change in paleocurrent direction occurred. On the Manas section (S-Junggar), the main direction is towards the north and in the Kuqa sub-basin (N-Tarim), the measured directions globally indicate southward transport. This confirms the existence of a Mesozoic paleo-relief in the Tian Shan area (Hendrix et al., 1992). Furthermore, the growing Bogda Shan is eroded and drained towards the north in the Junggar basin and to the south in the Turfan basin.

Cretaceous

The Lower Cretaceous depositional environments are mostly constituted of fan delta on the basin margins and of lacustrine systems in the Junggar Basin center. The depositional environment in the Turfan basin is still unclear (Table 2). Shao et al. (1999) interpreted lacustrine deposits from the northern and middle parts of the basin (one studied section at Lianmuqin, Fig. 2.1b), while Hendrix et al. (1992) first described highly aggradational braided fluvial systems from the Flaming Mountains Section, and then well oxygenated lacustrine deposition (Hendrix et al., 2000). In the Kuqa sub-basin (N-Tarim) lacustrine and delta deposits are described by Wang et al. (2013), while Hendrix et al. (1992) do not describe any lacustrine, but braided fluvial deposits and later (Hendrix et al., 2000) poorly oxygenated, local lacustrine and alluvial deposition in the Lower Cretaceous series. Hendrix et al. (1992) observed an unconformity between the Lower and Upper Cretaceous deposits in the S-Junggar, Turfan and N-Tarim (Table 2).

In the S-Junggar, Turfan and N-Tarim basins, the Upper Cretaceous series consist in braided fan-delta, braided fluvial to alluvial deposits. Hendrix et al. (1992) and Shao et al. (1999) observed an unconformity at the Cretaceous-Paleocene boundary in the Turfan Basin.

Paleocurrent directions are globally similar to the Jurassic ones in the S-Junggar and N-Tarim basin (e.g. Hendrix et al., 1992; Li et al., 2004; Bian et al., 2010). No data are available for the Turfan Basin.

The Mesozoic sediment deposits in the Tian Shan area record three general prograding phases, each evolving from a lacustrine to an alluvial fan environment: one during the Triassic, one during the Jurassic and one during the Cretaceous.

Paleocene

The S-Junggar depositional environments consist in braid-delta and shallow lakes in the Paleocene. Braided fluvial and alluvial to lacustrine environments, locally with salt-water-lakes in arid climate (evaporates) have been described in the Turfan Basin.

3) CLIMATE

The Lower to early Middle Triassic is dominated by arid climatic conditions in the Tian Shan area, as shown by paleosols (Li et al., 2004; Thomas et al., 2011) (Table 2). From late Middle to Late Triassic a change to warm and wet conditions has been observed throughout the three basins (Junggar, Turfan and Tarim) based on sporopollen assemblages (e.g. Shao et al., 1999; Li et al., 2004; Ashraf et al., 2010; Pan et al., 2013).

The Late Triassic and Early Jurassic periods correspond to a maximum of humidity, as shown by the formation of thick coal-seams all over Central Asia and the development of forests preserved as fossil wood (e.g. Hendrix et al., 1992; Shao et al., 2003; Le Heron et al., 2008).

The Lower and Middle Jurassic seem to be still warm and moist periods, but also mark the onset of the development of dry seasons at the end of the Lower Jurassic (e.g. Eberth et al., 2001; Shao et al., 2003; Ashraf et al., 2010; Sha et al., 2011; Pan et al., 2013). From Bathonian time, sedimentological and palynological data show the first basin-wide evidence of seasonally dry conditions in the Junggar basin (Eberth et al., 2001; Ashraf et al., 2010). During the Late Jurassic, a clear change from coal deposition to redbeds (Qigu Formation) is observed all over northwest-China and indicates a “strong” aridification leading to a semi-arid to arid climate (e.g. Allen et al., 1991b; Hendrix et al., 1992; Shao et al., 2003; Li et al., 2004; Eberth et al., 2001; Pan et al., 2013). It can also be interpreted as “strong” monsoonal conditions with a prominent dry season (Ashraf et al. (2010). Hendrix et al. (1992) suggested that this system stopped during the Middle Jurassic. Climatic

conditions in the area, as well as in the whole Central Asia are warm and seasonally arid from the end of Lower Cretaceous (e.g. Eberth et al., 2001; Li et al., 2004; Jiang et al., 2008). During the Cenozoic, strong seasonality prevailed with arid to semi-arid conditions during the summer, as shown during the Middle Eocene (e.g. Sun and Wang, 2005; Bougeois et al., 2014). In the Junggar Basin a Mid-Miocene climate optimum could be inferred from palynological data (Sun and Zhang, 2008). From the Early Miocene (Ramstein et al., 1997) or even earlier from the Eocene (Huber and Goldner, 2011), modellings show that the onset of the East-Asian monsoon modified humidity distribution in Asia and probably increased the aridity in Central Asia.

SEDIMENTOLOGICAL ANALYSES

We performed a detailed sedimentological study on six sections (the Manas, northern Totoun, southern Totoun, Wusu, Nileke, Bayanbulak and Yaha sections) located in the northern and southern foothills of the Tian Shan range, as well as in two intramontane basins (Fig. 2.1). Most of the outcrops are situated in river incision valleys or canyons, and their neighborhood. Each section is called after the name of the adjacent river valley.

For these six sections, a sedimentological analysis was performed using a standard faciological method. Facies were defined on the basis of their lithology, sedimentary structures and biological remains before to be interpreted in terms of depositional processes. In the studied sedimentary series, these facies co-occur in several facies associations, which were interpreted in terms of depositional environments. Those facies associations, which appear in specific locations along the studied sections, were used to divide the latter in several units with different environmental characteristics. Palaeocurrent data were also included to our sedimentological analysis in order to constrain local paleoflows and the drainage pattern in the Tian Shan area. Finally, we determined the clast composition of some conglomerates, in order to constrain sediment provenance and catchment area evolutions.

1) Facies model and depositional environments

In the studied sections, we distinguished 28 sedimentary facies (Table 3 and Fig. 2.9) related to eleven main depositional environments going from lakes, to deltas, alluvial plains and alluvial fans

(Table 4). The interpretation of these sedimentary facies is based on references from the literature and are summarised in Table 3.

A: Lake deposits

Five facies can typically compose the faciological units linked to lacustrine environments (L1 and L2) (Table 3 and 4, Fig. 2.2.1, 2.2.2): one carbonaceous facies (facies OM), one calcareous facies (Cal1), one silty facies (Sub1), one sandy facies (Wav1) and one rare sandy to conglomeratic facies (SGr1). Among these facies, the thick-bedded (centimeters to meters) siltstone layers (Sub1), which usually dominate the lake deposits, as well as the thin-bedded (centimetres to decimetres) silty carbonate to calcareous siltstone or sandstone beds (Cal1), indicate respectively terrigenous and calcareous, long-lived, subaqueous sediment fallout. In addition, the decimetric coal layers indicate anoxic organic-matter accumulation (OM). Some centimetric to decimetric sandstone beds with symmetrical ripples (Wav1) also point out wave deposition or reworking. Finally, a few centimetric to decimetric sandstone to granule layers exhibits sequences of normal gradings and ripples to mega-ripples (SGr1) that suggest the occurrence of turbiditic currents. All these deposits are often of dark grey or black colour because of a high content of organic matter. However, they can also be blue- or green-, and sometimes red-coloured. Bioturbation (e.g. horizontal burrows such as *Planolites*), which is a typical feature for these facies, can be faint to intense.

The deposition depth of these sediments must range several meters to several tenths of meters, but remains difficult to estimate. Nevertheless, we can differentiate distal from more proximal lacustrine environments because of their relative proportion of episodic clastic events, such as the gravitary deposits (SGr1), and the occurrence of wave ripples (Wav1). Distal lacustrine environments (L1) are characterized by few small (centimeters to decimeters) sandy to gravelly gravitary sequences caused by turbiditic currents (Table 4). Locally, some massive (centimeters to decimeters) beds of calcareous mud- and siltstones (Cal1) also occur. By getting more proximal and shallower, lacustrine environments (L2) show either an increase in number and a thickening of the gravitary sequences, or small-scale (height of a few centimeters) symmetrical ripples in thin (centimeter to decimeter) wave-influenced sandy layers (Wav1), which are typical of shallow water bodies (Table 4).

B: Delta deposits

19 facies can compose the faciological units linked to deltaic environments (D1, D2 and D3) (Table 3 and 4): one carbonaceous facies (OM), one calcareous facies (Cal1), two silty facies (Sub1, Sub2), two heterolithic facies (SFI1, ASt2), nine sandy facies (Wav1, SGr2, SGr3, SFI2, AFI1, ASt1, ASt3, ASt4, ASt5), one sandy to pebbly facies (SGr1), one conglomeratic facies (ASt6) and two pedogenic facies (Ped1 and Ped2).

Nine of these facies correspond to subaqueous deposits. Five of these subaqueous facies also occur in lacustrine environments (OM, Cal1, Sub1, Wav1 and SGr1) (see part A: Lake deposits). The four other ones are typical of deltaic units (SGr2, SGr3, SFI1 and SFI2) (Fig. 2.2.1, 2.2.2). The sandstone beds with soft-sediment deformations (SGr2) correspond to slump deposits. The thick-bedded (decimeters to meters) homolithic sandstone layers with basal flute casts and dewatering structures (SGr3) indicate flood-related gravity currents. The thick-bedded (centimeter to meters) heterolithic siltstone and fine sandstone beds (SFI1), locally with horizontal or asymmetrical laminations, suggest the occurrence of deposition from subaqueous gravity flows alternating with sediment fallout. The thick-bedded (decimeters to meters) homolithic sandstone layers (SFI2), locally with mega trough cross-bedding or horizontal planar bedding, point out deposition of a sandy bedload transported by subaerial to subaqueous unconfined flows on mouthbars. All these deposits are often of dark grey or black colour because of a high content of organic matter. However, they can be locally blue- or green-, and sometimes red-coloured. Locally, some coal occurs as reworked material in the siltstone and sandstone facies. Light to strong bioturbation (burrows), mud rip-off clasts and plants fragments are also common features of all these deposits.

The remaining ten facies observed in deltaic units (Sub2, AFI1, ASt1, ASt2, ASt3, ASt4, ASt5, ASt6, Ped1 and Ped2) correspond to subaerial deposits. These facies can also occur in alluvial environments (see part C: Alluvial plain deposits). The thin-bedded (several centimetres to several decimeters) homolithic siltstones (Sub2), can be associated to organic matter (OM), or to root casts (Ped1). They indicate subaqueous sediment fallout in a thin body of water close to subaerial deposition, locally with soil-formation in a wet environment. Thin (centimeters to decimeters) planar laminated siltstones (Sub2) can also present mudcracks or calcareous nodules (Ped2) that stand for wet-dry pedogenic processes. The relatively thick-bedded (several centimeters to several decimeters) homolithic sandstone beds with climbing ripples (AFI1) indicate deposition of a sandy bedload transported by a subaerial sheet flows corresponding probably to overbank. The thick-bedded (decimetre to several decimeters) sandstone beds with asymmetrical ripples (ASt1) formed due to the deposition of a sandy bedload transported by subaerial stream or sheet flows. They can also

occur as heterolithic deposits (ASt2) resulting from multiple stacked deposition of a sandy bedload transported by subaerial stream or sheet flows. Thick-bedded (several decimetres to several metres) sandstones with planar (ASt5), through cross-bedding (ASt3) or both (ASt4) result from deposition of a sandy bedload transported by subaerial stream-flows. Clast-supported, pebble conglomerates with mega-ripples, locally matrix-rich (sandy) (ASt6) point out deposition of a sandy to gravelly bedload transported by subaerial stream-flows. The color of these deposits varies from grey-beige and dark, probably when deposited under rather humid conditions, to red, purple and orange when deposited under warmer and drier conditions.

On the basis of the association of all these facies occurring in various proportions, three deltaic environments can be distinguished, going from prodelta to front delta and deltaic plain (Table 4). The prodelta environments (D1) resemble to the proximal lake ones, but show a higher proportion of flood-related gravitary events (SGr1, SGr2, SGr3, SFl1). The front delta environments (D2) consist mostly of homolitic, generally well sorted, stacked (over meters to tenths of meters) sandstone layers that correspond to mouth bar deposits (SFl2). Slumping (SGr2) is also common in these thick deposits. Some of these facies go laterally into heterolithic muddy to sandy siltstones (Sub1), locally bioturbated (undetermined burrows) and locally organic matter-rich (OM). In front delta environments, erosional basal boundaries are rare and there is no channel deposit. The deltaic plain environments (D3) are at the interface of different processes (flooding, sediment fallout, wave-influenced processes). These environments resemble to distal alluvial plains (see part C: Alluvial plain deposits), but they are influenced by a nearby permanent body of standing water. There are dominated by thick (decimeter to meters) sandstones to fine-grained conglomerate facies (ASt3, ASt4, ASt5, ASt6), which are deposited in distributary channels. Climbing and asymmetrical ripples in fine sandstones (AFI1, ASt1) associated to thin (several centimetres to several decimeters) silty to fine sand suspension-fallout deposits (Sub2) are typical for overbank sediments. However, some sandstone beds with symmetrical ripples (Wav1) and faint to strong bioturbation show that the deltaic plains can be temporarily under water. On the other hand, root casts (Ped1) and calcareous nodules (Ped2) imply that deltaic plains are also temporarily subaerial and that sedimentation can stop to leave room for pedogenic processes to occur.

C: Alluvial plain deposits

14 facies can compose the faciological units linked to alluvial plain environments (AP1, AP2 and AP3) (Table 3 and 4): one carbonaceous facies (OM), one silty facies (Sub2), one heterolithic facies (ASt2), eight sandy facies (AFI1, ASt1, ASt3, ASt4, ASt5, Eol1, Eol2, Eol3), one conglomeratic facies (ASt6) and two pedogenic facies (Ped1 and Ped2).

11 of these facies also occur in lake (OM) and deltaic plain environments (OM, Sub2, AFI1, ASt1, ASt2, ASt3, ASt4, ASt5, ASt6, Ped1 and Ped2) (see part A: Lake deposits and part B: Delta deposits). In these deposits, mud rip-up clasts are common at the base or throughout the whole beds (ASt2, ASt3 and ASt4). Well preserved plant fragments can also occur, often at the bottom of the strata (OM). Locally trees in standing position attest of emerged deposits, as well as root traces and rootlets (Ped1). However, except the latters, bioturbation is often very limited. Among these facies, the coal layers and organic matter rich mud (OM) are evidences for deposition in wet environments, whereas calcareous nodules from poorly to highly coalescent (Ped2) are typical for more semi-arid environments. The color of these deposits also varies from grey-beige and dark, probably when deposited under rather humid conditions, to red, purple and orange when deposited under warmer and drier conditions.

The other three only occur in alluvial plain environments (Eol1, Eol2, Eol3) (Fig. 2.2.3). The thick-bedded (several decimetres to decametres) sandstones with sandflow and grainfall laminations, as well as inverse climbing ripples (Eol1) are typical of migration of aeolian dune deposits. The thick-bedded (several decimetres to decametres) sandstones with very low angle cross-laminations and inverse climbing ripples (Eol2) point out deposition of aeolian sand sheets during high wind velocities. The thin-bedded (decimetres) sandstones with preserved wind ripples, adhesion structures and lack of inclined forest (Eol3) indicate either deposition in interdunes alongside the dunes, or deposition of sand transported by aeolian sand sheets within a wet environment.

The criteria used to differentiate the distal from the more proximal alluvial plain environments are the grain size of the channel sediments and the dominant type and size of their beddings (e.g., cross- or horizontal-beddings) and associated bedforms (e.g., mega-ripples or bedload sheets), together with the abundance of the overbank deposits and the frequency of some particular processes (e.g., aeolian). Distal alluvial plain environments (AP1) consist of sandy channel deposits (ASt3, ASt4 and ASt5) associated to well-developed overbank sediments (Sub2, AFI1, ASt1 and ASt2) with some coal layers (OM) or few pedogenic evidences (Ped1, Ped2). In some units, stacked aeolian deposits (Eol1, Eol2, Eol3) are also characteristic of distal alluvial plains and indicate semi-arid to arid environments. Median alluvial plain environments (AP2) present sandy to pebble conglomerate

channel sediments (ASt3, ASt4, ASt5 and ASt6), overbank deposits comparable to the ones associated to distal alluvial plains (Sub2, AF1, ASt1 and ASt2 with OM or Ped1, Ped2), but in a smaller proportion. Finally, the proximal alluvial plain environments (AP3) contain numerous pebble conglomerate channel sediments (ASt6) associated with some overbank deposits. In some units, the sandy and pebbly channel deposits are also interbedded with aeolian deposits (Eol1, Eol2, Eol3) that are characteristic of median or proximal, semi-arid to arid alluvial plains.

D: Alluvial Fan deposits

Four pebbly facies compose the faciological units linked to alluvial fan environments (AF1, AF2 and AF3) (Table 3 and 4): ASt7, ASt8, AGr1, AGr2.

Four facies are typical of alluvial fan units (ASt7, ASt8, AGr1, AGr2) (Fig. 2.2.5). The thick-bedded (metres to tens of metres) clast-supported pebble to cobble conglomerates (ASt7) point out deposition of a gravelly bedload transported by subaerial stream- or sheet-flows. Locally, the same kind of deposits also show winnowed surfaces or root casts (ASt8, Ped1), which indicate that these stream- or sheet-flows can have been ephemeral. On the other hand, the thick-bedded (several decimetres to several metres) matrix-supported pebble to cobble conglomerates with a sandy matrix (AGr2) are the result of subaerial debris flows. The thick-bedded (metre to metres) clast- to matrix-supported pebble to cobble conglomerates with an upward increase in matrix content (AGr1) indicate that they have been transported by a subaerial debris flow entering into a body of standing water. All these facies can be grey or red-brown coloured.

The criteria used to differentiate the distal from the more proximal alluvial fan environments are the grain size of the channel sediments and the type and size of their beddings (e.g., cross- or horizontal beddings) and associated bedforms (e.g., mega-ripples, sours and fills or bedload sheets), as well as the relative proportion of the sandy and conglomeratic facies and stream-flows, sheet-flows and gravitary events. Indeed, distal alluvial fan environments (AF1) consist of pebble conglomerate channel deposits (ASt7 and ASt8) associated to fine-grained interbeds (Sub2, AF1, ASt1). Large-scale mega-ripples and bedload sheets can be observed in both conglomerates and sandstones. Median alluvial fan deposits (AF2) resemble to the distal ones, but with less large-scale bedforms and fine-grained interbeds. Locally, debris flows can occur within distal and median alluvial fan environments, but remain limited. In proximal alluvial fan environments (AF3), debris flows (AGr1 and AGr2) are most common and represent thick events. They are associated with thick pebble to

cobble conglomerates (ASt7 and ASt8) with numerous bedload sheets and a few mega-ripples. In some distal, median and proximal alluvial fan conglomerates (ASt8), thin (decimeters) winnowed surfaces indicate ephemeral stream- or sheet-flows that could correspond to flashflood events under semi-arid to arid conditions.

2) Vertical evolution of depositional environments: section analyses

The sedimentological columns of the studied sections were usually measured at a resolution of one cm representing ten metres (1/1000), but they are presented here at a scale of one centimetre representing 100 metres (1/10 000).

The associated palaeocurrent directions were derived from imbricated clasts and cross-beddings or deduced from strikes of channel axes (no flow direction, only axe of the current). The dip of the studied series usually exceeds 10° and requires a correction of the measured palaeocurrent directions. In addition, the clast compositions of some conglomerates were evaluated by means of 20 cm x 20 cm squares with a minimum of 300 clasts counted on each station. Clast lithologies were grouped into four categories: the felsic, mafic, calcareous and terrigenous rocks. The felsic category includes granites, rhyolites and felsic metamorphic rocks. The mafic one comprises basalts, andesites and volcano-sedimentary rocks. The calcareous category groups limestones. Below, each section is described from bottom to top.

2.1) Northern foothills

2.1.a) The Manas Section

The Manas Section is our longest section (ca. 8100 m) (Fig. 2.3) spanning from Upper Triassic to Paleogene ages according to the geological map (XBGMR, 1978a). It is the reference section of this work. The section is situated on the northern flank of the Tian Shan Range (Fig. 2.1.b), about 50 km south of the city of Shihezi and about 130 km from both the Wusu Section (in the west) and the Totoun Section (in the east) that will be discussed below. It is divided into two parts: the stratigraphic lower part is situated east of the Ningjia river and represents ca. 2550 m of deposits (base of the section: GPS 43°54'3.84"N/ 85°39'46.20"E, top of the section: GPS 43°55'31.14"N/ 85°44'26.58"E).

The second part of the section is situated west of the Manas river and comprises 5550 m of deposits (base of the section, GPS 43°54'6.70"N / 85°51'7.70"E, top of the section: GPS 43°58'12.72"N/ 85°48'42.60"E). We correlated these two sections with the help of the geological map (XBGMR, 1978a) and of satellite images (Google Earth).

34 faciological units were distinguished, which are interpreted as various depositional environments from distal lake (L1, e.g. unit M31) to proximal alluvial fan (AF3, e.g. units M4, M29, M32) (Fig. 2.3, Table 3, 4 & 5). The first unit M1 consists of alternations of thick sandstone bars with siltstone interbars, locally with debris flow deposits (Fig. 2.10.1.a). Unit M1 was deposited in the proximal part of a median alluvial plain environment (AP2). M2 is dominated by clast-supported, thick conglomerates with rounded pebble clasts, typical of a median-distal alluvial fan environment (AF2-AF1). It is followed by a short retrogradation to unit M3 that is an outcrop gap but seems to show deposition in an alluvial plain (AP). A new progradation leads to unit M4 that consists of a clast-supported conglomeratic bar, deposited in a proximal alluvial fan (AF3) (Fig. 2.10.1.b). A progressive retrograding trend starts with unit M4 and goes to unit M5 that points out to some environments ranging from a median to distal alluvial fan (AF2-AF1). This trend ends with unit M6 that seems to be dominated by thick fine sandstone deposits, probably from a distal alluvial plain environment (AP1). After a short progradation, unit M7 was deposited in an alluvial fan environment with some debris-flow transport (AF). The section continues in a parallel valley, situated east of the previous one. The prograding trend ends with unit M8 that was deposited in a median alluvial fan deposits (AF2). It is then followed by a retrogradation trend from units M8 to M10. Unit M9 was deposited in a median alluvial plain environment (AP2) (Fig. 2.10.1.c). Unit M10 is hardly exposed, but seems to contain very thick silts and fine sandstones with thick sandstone bars showing progradation (Fig. 2.10.1.d). This unit was deposited in a prodelta environment (D1) and indicates a prograding trend going from the proximal lake (L2) (at the base) to the distal frontdelta (D2) (at the top). This progradation continues to unit M11 deposited in a median alluvial plain (AP2) (Fig. 2.10.1.e). It is then followed by a retrogradation from units M11 to M13. Unit M12 is dominated by stacked thick sandstone deposits, typical for channels in median-distal alluvial plain environments (AP2-AP1). The large outcrop gap spanning units M13 and M14 seems to show a prograding trend. Unit M13 was deposited in a deltaic plain environment. Unit M14 is an alternation of clast-supported conglomerates, sandstone beds and locally organic matter-rich layers. It was deposited in a proximal-median alluvial plain (AP3-AP2). After this a steady retrogradation goes from unit M14 to unit M18. First unit M15 was deposited in a distal alluvial plain environment (AP1) (Fig. 2.10.1.f), followed by unit M16 that is not continuously exposed but presents typical sediments of a deltaic plain environment (D3), such as coal deposits (Fig. 2.10.2.a). The second part of the Manas section starts at the end of unit M16.

Unit M17 consist of alternation of dark organic matter-rich silty deposits and sandstones, typical for a frontdelta environment (D2). The retrograding trend is immediately overlaid by a progradation trend from units M18 to M22. Unit M18 is first composed of thinning-out sandstone bars followed by thick dark organic matter-rich siltstones. These deposits shift to thickening-out sandstones bars alternating with thick dark and organic matter-rich siltstones and are typical for a prodelta (D1). Unit M19 was deposited in a frontdelta (D2), with some trends to the prodelta (D1) (Fig. 2.10.2.b). They are exposed in the canyon of the Manas River and the boundaries between our units were not accessible (Fig. 2.10.2.c). This is the reason why we could not determine precisely the individual thickness of these three units. Unit M20 contains alternation of organic matter and plant remains-rich siltstones and sandstone beds with thinning- and thickening-out sequences, typical of a deltaic plain. Unit M21 shows distal alluvial plain deposits (AP1). Unit M22 ends the progradation trend and is composed of several small prograding sequences (ca. 10 m-thick) that end with a thick sandstone bar. Unit M22 was deposited in a distal-median alluvial plain (Fig. 2.10.2.d). The trend of depositional environment going from units M22 to M24 is retrograding. The shift from unit M22 to unit M23 is abrupt. Unit M23 shows several meters to tens of meters-thick progradation sequences and was deposited in a frontdelta environment (D2). Unit M24 was deposited in a prodelta (D1) (Fig. 2.10.2.e) and is separated from unit M25 by an abrupt shift and a short progradation. Unit M25 was deposited in a distal alluvial plain (AP1) (Fig. 2.10.3.a). It is followed by a short retrogradation to unit M26 that was deposited in a very distal alluvial plain (AP1), on the edge to a deltaic plain. A new prograding trend begins with unit M26 and goes to unit M27 that is characteristic of a distal alluvial plain environment (AP1) (Fig. 2.10.3.b). This prograding trend continues to unit M28 after a relatively sharp shift. This unit is indicative of deposition in the most distal part of a distal alluvial fan environment (AF1) (Fig. 2.10.3.b, c, d). The prograding episode ends with unit M29 that consists of typical deposits of a proximal alluvial fan (AF3) (Fig. 2.10.3.b, c). The transition to the next unit (M30) is very sudden, and shows a sharp retrogradation from the most proximal environment (AF3) to proximal lake environment (L2). A retrogradation trend spans units M30 and M31, where sediments were deposited in a distal lake environment (L1) (Fig. 2.10.3.e). This unit is separated from unit M32 by a gap of 430 m. A sharp shift separates units M31 and M32 underlined by a progradational trend. Unit M32 was deposited in the most proximal part of a median alluvial fan (AF2) (Fig. 2.10.3.f and 2.10.4.a), locally going up to the proximal alluvial fan (Fig. 2.10.4.b). The last eight meters of unit M32 are characterized by a white continuous level of neo-formed calcareous impregnation of the matrix, also called calcrete (facies Ped3, Table 3; Fig. 2.10.4.c). This impregnation is parallel to the bedding and laterally continuous over tens of kilometers (Fig. 2.2.5, 2.3, 2.18; Chapter 4 of this thesis). A retrograding trend follows with unit M33, which is characteristic of a median alluvial plain environment (AP2) (Fig. 2.10.4.d) and unit M34 that was deposited in a proximal lake (L2).

Ages and complementary data of the Manas series

Units M1 and M2 are supposed to be of Triassic age (XBGMR, 1978a). The boundary between Triassic and Jurassic deposits is situated within unit M3. The boundary separating the Lower from the Middle Jurassic is contained in unit M13, the one separating the Middle from the Upper Jurassic is in unit M23. In our area the boundary between Jurassic and Cretaceous deposits has been interpreted in different ways and is probably contained in the Kalazha Formation (unit M29) (Eberth et al., 2001). The shift from Lower to Upper Cretaceous is at the boundary between units M31 and M32. Units M33 and M34 overlay the calcrete and are Cenozoic deposits.

Paleocurrent directions measured in units M1 and M2 are spread, globally indicating an east-west direction, while measured directions from unit M3 upwards are spread around the north-south direction. This change in paleocurrent directions seems to occur at the Triassic – Jurassic boundary.

Five stations have been chosen to determine the clast composition of conglomerates: at the base of the Lower Jurassic deposits, at the base of the Middle Jurassic, at the base and top of the Kalazha Formation (late Upper Jurassic) and just below the calcrete (supposed of late Upper Cretaceous age). Clast composition shows a high increase in terrigenous rock proportion in the Middle Jurassic (from 27 to 76 %), reaching up to 96 % at the top of the Kalazha Formation. From the base of the latter formation, we note the appearance of calcareous rocks, increasing from 1 % up to 4 %. In the late Upper Cretaceous, the proportion of felsic and mafic rocks are in equilibrium with terrigenous rocks. Felsic and mafic rocks are eroded from the Paleozoic basement of the Tian Shan Range. In contrast, the terrigenous category gathers sandstones and shales, which correspond to the Meso-Cenozoic cover of the range. This evolution suggests sediment recycling/cannibalism through Jurassic and Cretaceous period, followed by renewed basement erosion during Cretaceous, due to complete erosion of the sedimentary cover.

Units M1 to M24 show deposition in a rather wet environment. From unit M25 upwards, calcareous nodules and others features, root traces in silty interbars) indicate deposition in a semi-arid environment.

2.1.b) Totoun River area

The area we visited close to the Totoun River is situated on the northern flank of the Chinese Tian Shan, about 115 km east of the Manas section (Fig. 2.1.b). The first section (South Totoun section) is situated south of the Totoun dam and is 1983 m thick (base of the southern part of the section: GPS 43°41'49.10"N/ 87°11'33.90"E, top of the southern part of the section: GPS 43°43'6.80"N/ 87° 8'34.90"E). The second section (North Totoun section), north of the dam, presents 2505 m of deposits (base of the northern part of the section: GPS 43°46'24.30"N/ 87°15'8.16"E, top of the northern part of the section: 43°48'28.38"N/ 87°17'3.54"E). The two measured sections range in age from Mid-Jurassic to Tertiary.

The southern and northern sections can be divided respectively in ten and nine faciological units (Fig. 2.4, 2.5; Table 3, 4 & 5).

2.1.b.1) Southern Totoun Section

The first unit, TS1 consists of relatively stacked sandbars that can be thickening-out and coarsening-up, with silty and fine sandstone interbars. Channel deposits occur with typical erosional basal boundaries and lateral thickness variations and are characteristic of a distal-median alluvial plain environment (AP1-AP2). The transition to unit TS2 is abrupt and corresponds to a retrogradation. A prograding episode starts with unit TS2 that was deposited in a delta (D2 to D1, and D3). This trend continues with unit TS3 that comprises commonly bioturbated (pedogenic) silt- and sandstone heterolitics, locally with some calcareous nodules. It is typical of the deposition in a distal alluvial plain environment (AP1). A short retrogradation follows and leads to unit TS4 that was deposited in a deltaic plain (D3). A new prograding trend starts in unit TS4 and continues with unit TS5 that shows typical aeolian deposits from a distal-median alluvial plain (AP1-AP2). The progradation continues to unit TS6 that were deposited in a median-distal alluvial plain (AP2-AP1). The prograding trend ends after a sharp shift to unit TS7 that is characteristic of deposition in a median alluvial fan environment (AF2). It is followed by a short retrogradation to unit TS8 that shows several stacked sand bars and multiple stacked coarsening-up and fining-up sequences draped with centimeter-thick lags of pebbles. Unit TS8 was deposited in a median-distal alluvial plain (AP2-AP1). A new sharp shift and progradation trend occurs between units TS8 and TS9. Unit TS9 is typical of a median alluvial fan environment (AF2), with a tendency to the proximal part of it. Unit TS10 is the last unit of this section and was deposited after a sharp shift. Unit TS10 shows the last retrograding trend and goes from a prodelta (D1) to a proximal lake environment (L2).

Ages and complementary data of the South Totoun series

Unit TS1 is supposed to be of Middle Jurassic age (Totounhe Formation) (XBGMR, 1978b; Fig. 2.4). The boundary between the Middle and Upper Jurassic is situated within unit TS2. Unit TS3, TS4, TS5 and TS6 belong to the Upper Jurassic Qigu Formation and units TS7, TS8 and TS9 to the Kalazha Formation. The boundary separating the Upper Jurassic from the Lower Cretaceous series corresponds to the limit between units TS9 and TS10.

Three paleocurrent directions have been measured in units TS7, TS8 and TS9, and indicate the same direction towards the north (Fig. 2.4).

The first unit (TS1) seems to be deposited in a wet environment. The general progradation goes together with an aridification trend as supported by the aeolian deposits. A semi-arid climate seems to have dominated from unit TS3 to TS9, with an excursion at unit TS4 where deposition in a standing body of water did not provide any information about the climatic conditions. The same issue arises in unit TS2 that does not record the start of the aridification, because sediments are deposited under water.

2.1.b.2) Northern Totoun Section

Unit TN1 corresponds to alternation of thick massive beige sandbars with erosive basal boundaries and lateral thickness variations, and red heterolitics characteristic of prodelta deposits (D1) that give rise to some stacked thick sandbars, typical of front-deltaic facies (D3). This progradation continues to unit TN2 deposited in a median (locally distal) alluvial plain environment (AP2). The transition to unit TN3 shows a short retrogradation. Above this, unit TN3 was deposited in a deltaic plain environment (D3). A new short progradation takes place before the deposition of the first aeolian deposits (unit TN4) in a median-distal alluvial plain environment (AP2-AP1). The boundary to unit TN5 is an erosive unconformity. Most of unit TN5 is difficult to access because of steep cliffs. However the large structures in the landscape and small outcrops in the river canyon enabled us to interpret aeolian deposition of dunes and sandsheets with wet interdunes in a distal alluvial plain environment (AP1). The overlaying unit TN6 starts probably earlier in the serie, but the boundary is not accessible. Unit TN6 was probably deposited in a median-distal alluvial plain environment (AP2-AP1). A new retrograding trend starts at the end of this unit and shows a sharp shift to unit TN7 that was deposited in a proximal lake (L2), close to the prodelta environment (D1).

Ages and complementary data of the North Totoun series

Units TN1 to TN4 are considered to be part of the Upper Jurassic Qigu Formation and units TN5 and TN6 of the Upper Jurassic Kalazha Formation. Unit TN7 is probably of Cretaceous age (Fig. 2.5).

At the base of the northern Totoun section, two main paleocurrent directions could be observed in units TN1 and TN2: a flow-direction to the west (in-between N257 – N292, $n = 6$) and one to the east-east-north (N060-N075, $n = 4$) (Fig. 2.5).

Units TN1 to TN3 were deposited under water, therefore they do not provide more details about the onset of the following aridification. Unit TN4 contains some evidences for a semi-arid climate such as mudcracks. Unit TN5 shows aeolian deposits, that indicates a semi-arid to arid climate.

As a result, we observed that the South Totoun section seems to have been situated in a slightly more proximal position than the North Totoun section in general, especially when comparing the aeolian units.

2.1.c) Wusu Section

The Wusu section (1465 m in total) is located 115 km west of the Manas section (Fig. 2.1.b) and is called after the closest town. According to the geological map this section spans Middle to Upper Jurassic ages (XBGMR, 1973a). It is divided into two parts. The stratigraphic lower part, situated west of Saili Ketu village comprises 982 m (base of the section: GPS 44° 8'1.86"N/ 84°27'46.80"E, top of the section: GPS 44° 9'1.68"N/ 84°26'53.82"E). We expect that this first part overlaps the second part of the section, situated east of Saili Ketu village and comprising 483 m (base of the section, GPS 44° 8'10.32"N/ 84°30'18.42"E, top of the section: GPS 44° 8'43.92"N/ 84°30'12.48"E). The continuity of the serie between the two sections could not be determined at the scale of decimeters-thick bars and we expect some overlap between the two parts of this section.

Eleven faciological units on the western and lower parts, and six units on the eastern and upper part of the section can be identified (Fig. 2.6, Table 3, 4 & 5).

Lower section

The first unit, W1 was deposited in a proximal-median alluvial fan environment (AF3-AF2). It shows a retrograding trend to unit W2 that was formed in a proximal alluvial plain environment (AP3). The shift to unit W3 is sharp and shows a progradation trend to a proximal-median alluvial fan environment (AF3-AF2). This is followed by a very sharp change and retrogradation to unit W4. The depositional environment of unit W4 is a distal lake (L1) (in the most proximal part of it). From units W4 to W7 the general trend is prograding. Unit W5 was deposited in a frontdelta (D2) (locally on the edge to the deltaic plain or the prodelta) and unit W6 in a proximal-median alluvial plain environment (AP3-AP2). This trend ends with unit W7 which is typical of a proximal-median alluvial fan environment (AF3-AF2), locally going to an alluvial plain environment. It is then followed by a short retrogradation to unit W8 that shows alternation of thick fine to coarse sandstone layers with few conglomeratic lenticular bodies and locally with tuffs layers (Tuf1), often interbedded with coal deposits or organic matter-rich silt- and sandstones. These deposits are characteristic of the most distal part of the proximal alluvial plain environment (AP3). Unit W8 shifts to unit W9 after a sharp progradation. Unit W9 consists of typical deposits of a proximal alluvial fan (AF3), sometimes closer to the median part of the alluvial fan. It is followed by a retrograding trend from units W9 to W11. Unit W10 was deposited in a distal alluvial fan environment (AF1) and unit W11 is poorly exposed, but was probably deposited in a distal alluvial plain (AP1). A new prograding episode starts with unit W11 and leads to unit W12 which is typical of a median alluvial plain environment (AP2). This trend ends within unit W12 and shifts to a retrograding trend to unit W13 which is supposed to be deposited in a median to distal alluvial plain environment (AP2-AP1). This is followed by a short progradation to unit W14, which shows conglomeratic deposits, alternating with sandstones. This unit was probably deposited in a median alluvial plain environment (AP2).

Upper section

A retrograding trend encompasses units WE1 to WE3 and starts with unit WE1 that was deposited in a proximal alluvial plain (AP3) and shifts rapidly to unit WE2, after a very erosive boundary. Unit WE2 is characteristic of deposition in a median alluvial plain environment (AP2) and unit WE3 was formed in a slightly more distal median alluvial plain (AP2) than unit WE2. This is overlaid by a progradation until unit WE5 where unit WE4 is a short intermediary unit that was deposited in-between the distal alluvial fan (AF1) and the proximal alluvial plain (AP3). Unit WE5 was deposited in a median to

proximal alluvial fan environment (AF2-AF3). Thus the shift to unit WE5 is very sharp, as well as the one going with the retrogradation from unit WE5 to unit WE6. Unit WE6 was deposited in a median alluvial plain (AP2).

One should note that this section is characterized by numerous white tuff deposits (facies Tuf1) from units W8 to WE2.

Ages and complementary data of the Wusu series

We suppose Middle Jurassic units W1 to W9 to belong to the Xishanyao Formation and units W10 to W13 to the Totounhe Formation (Fig. 2.6). The boundary to the Upper Jurassic deposits is probably situated within units W14 or WE1, where the two parts of the section are overlapping. Units WE2 and WE3 can be associated to the Qigu Formation and units WE4 and WE5 to the Kalazha Formation. We suppose deposits of unit WE6 to be of Cretaceous age.

From unit W1 to W9, deposition probably occurred in a wet environment. There are no climate evidences in unit W12 and in the lower part of unit W13 except for some organic matter-rich layers that could indicate a rather wet environment. From unit W13 deposits are of various red, orange or purple color and could be deposited in a semi-arid environment, as attested by root traces and calcareous nodules from unit WE3.

2.2) Intramontane basins

2.2.a) The Nileke section

The Nileke section, together with the Bayanbulak section (see § 2.2b) are up to now the only two Mesozoic sections described in the smaller intramontane basins of the Chinese Tian Shan (base of the section: GPS 43°50'23.10"N/ 82°43'52.26"E, top of the section: GPS 43°49'35.70"N/ 82°44'1.32"E). The section is located in the Hexilagen Basin, on the southern flank of the North Tian Shan range (Fig. 2.1.b), ca. 250 km south of the Manas section. According to the geological map, this section of 910 m covers Middle Jurassic ages (XBGMR, 1973b). Formation names used here are similar to those in the northern foothills. The log stops just before a ca. 300 m-thick gap.

Six faciological units can be identified along this section (Fig. 2.7, Table 3, 4 & 5). The first unit, N1 corresponds to thick siltstone layers characteristic of prodelta deposits that give rise to some sandy front-deltaic facies (D1 to D2). This progradation continues to unit N2 deposited in a distal alluvial plain (AP1), locally in a lake. The transition to unit N3 is progressive and corresponds to a short retrogradation. Above, unit N3 points out to some environments ranging from deltaic plain to front delta (D3 to D2). A new prograding episode starts with unit N4, which is characteristic of a distal alluvial plain (AP1), and continues with unit N5, which is indicative of a median to slightly proximal alluvial plain (AP2 to slightly AP3). This prograding trend then ends with unit N6 corresponding to deposits of a median alluvial plain (AP2), a slightly more distal environment than the one of the previous unit N5. This section presents numerous decimeter- to meter-thick heterolithic coal layers or organic matter-rich silt- and sandstones.

Ages and complementary data of the Nileke series

This section is supposed to expose solely Middle Jurassic deposits (XBGMR, 1973b), probably of the Totounhe Formation (Fig. 2.7.1). Nevertheless, a single restricted outcrop of (Lower?) Cretaceous deposits are mentioned on the geological map (XBGMR, 1973b). Few meters of exposure show sediments probably deposited in a proximal alluvial plain environment (Fig. 2.7.2.a and b).

On the Nileke section, the observed sediments do not show any pedogenetic features and no red deposits, but organic matter-rich siltstones and sandstones, as well as coal layers. Many fossil plants are preserved in units N1, N2 and N3 (Fig. 2.7.1). We infer that most of the deposition occurred in rather wet environments.

2.2.b) Bayanbulak section

The Bayanbulak section is located in the Bayanbulak Basin, ca. 150 km south of the Manas section (base of the section: GPS 43° 4'10.38"N/ 84°17'13.20"E, top of the section: GPS 43° 4'35.19"N/ 84°16'49.93"E) (Fig. 2.1.b). The outcrop is located in a small canyon north of the basin. According to the geological map this 620 m thick section covers Middle to Upper Jurassic ages (XBGMR, 1969). Several north verging thrust faults have been described by Jolivet et al. (2010) and have been reported on the stratigraphic column.

Twelve faciological units can be identified along this section (Fig. 2.8; Table 3, 4 & 5). At the bottom of the section, units B1 and B2 consist of alternations of thick sandstone or conglomerate deposits, locally interbedded with siltstone and coal deposits. This is characteristic of deposition in a proximal alluvial plain environment (AP3), and unit B1 is slightly more distal than the overlying unit. This prograding trend continues in unit B3 dominated by thick conglomerates deposited in a proximal-median alluvial fan environment (AF3-AF2). Unit B4 is preceded and followed by sharp shifts and shows typical deposits of the very distal part of a median alluvial plain environment (AP2). The first shift stands for a retrogradation from unit B3 and the second for a progradation to unit B5. This unit was deposited in a proximal alluvial fan system (AF3). The transition to unit B6 is progressive and corresponds to a short retrogradation. Unit B6 was mostly deposited in the proximal part of a median alluvial fan (AF2) and was locally interbedded with proximal alluvial plain deposits (AP3). A new prograding episode starts with unit B7, which shows typical deposits of a median-proximal alluvial fan environment (AF2-AF3). We expect a fault zone to separate units B7 and B8 (Jolivet et al., 2010) and the thickness of the missing part of the section is unknown. Unit B8 contains deposits characteristic of a proximal alluvial fan (AF3). A sharp change between the most proximal and the most distal units of this section occurs between units B8 (AF3) and B9. Unit B9 was formed in the most distal part of a distal alluvial plain environment (AP1). A new progradation phase initiates from unit B9 to unit B11. Unit B10 is typical of the deposition in a median alluvial plain (AP2) and unit B11 shows sandstone-rich deposits in a median-proximal alluvial fan environment (AF2-AF3). This is overlaid by the last retrogradation episode from unit B11 to unit B12 that was deposited in a proximal-median alluvial plain environment (AP3-AP2). The last two units are separated by a sharp shift.

Ages and complementary data of the Bayanbulak series

Units B1 to B7 are supposed to be of Middle Jurassic age (XBGMR, 1969) (Fig. 2.8). The overlying units (B8 to B12) are possibly deposits of the Paleogene period.

Six stations have been used to determine the evolution of clast-compositions of conglomerates on the Bayanbulak section and are spread all over the section (Fig. 2.8). Felsic rocks are dominant and relatively constant (67% to 78%). The proportion of terrigenous rocks decreases after the first station, from 15% to 3 – 6 %. From station PET-B4, some calcareous rocks appear.

The top few meters of unit B8 are made of calcrete (facies Ped3, Table 3; see Heilbronn et al., in revision), which shows a CaCO₃-precipitation under semi-arid climatic conditions. It is the first unit

showing evidences of pedogenic processes that indicate deposition under a semi-arid climate, which will prevail in all following units.

2.3) Southern foothills

2.3.a) Yaha section

The Yaha section is the only one situated in the southern foothills of the Tian Shan range (Fig. 2.1.b), ca. 300 km south of the Manas section (base of the section: GPS 42°7'40.11"N/ 83°17'8.40"E, top of the section GPS 42° 4'52.60"N/ 83°15'48.30"E). According to the geological map this section of ca. 2800 m covers Lower Jurassic to Tertiary ages (XBGMR, 1970). In the field, some parts of the lithological column are doubled, separated by tectonic faults.

Twenty faciological units have been identified along this section (Fig. 2.9, Table 5). According to the geological map (XBGMR, 1970), nine are of Jurassic age and the following eleven units are of Cretaceous age. The first unit, Y1 is dominated by stacked, thick bedded sandstone deposits, associated to locally organic matter-rich siltstones and sandstones deposits that are characteristic of a well-developed front delta environment (D2). A prograding episode starts in unit Y1 and shifts to unit Y2, which shows deposits of a distal alluvial plain (or “swamp”) environment (AP1). It continues in unit Y3 that was deposited in a median alluvial plain (AP2), a slightly more proximal environment than the latter. Though unit Y3 is separated from Y4 by a major thrust fault zone, we expect the displacement to be mostly parallel to the layering. Based on lateral correlations we consider the loss of section to be insignificant for our study. A retrogradation occurs from unit Y3 to unit Y5. Unit Y4 is dominated by retrograding sequences and can be divided into two parts: first a deltaic plain (D3), overlaid by a front delta environment (D2). The top of unit Y4 is marked by a sharp shift and the retrogradation continues in unit Y5. Unit Y5 contains some massive calcareous mud- and siltstone or limestones bars in the upper part of the unit and was deposited in a lake environment caught between the proximal and distal parts (L2/L1). It shifts gradually into unit Y6 and shows a new prograding episode. Unit Y6 is supposed to be deposited in the most distal part of a proximal lake environment (L2). The prograding trend continues in unit Y7 that is typical of deposition in a deltaic plain environment (D3). It goes progressively into the heterolithic deposits of unit Y8 that were deposited in a median alluvial plain environment (AP2). The prograding trend continues to unit Y9 and is marked by a sharp shift between units Y8 and Y9. The thickness of the massive conglomerate (several decimeters to several tens of meters-thick) of unit Y9 varies considerably laterally and spans

30 to 50 m in average. The deposition environment of unit Y9 oscillates between the distal and proximal parts of the alluvial fan (AF1 to AF3). It is overlaid by a sharp shift that marks the boundary to unit Y10, which is an intermediate unit in the retrogradation from unit Y9 to Y12. Unit Y10 was deposited in a median alluvial plain environment (AP2), unit Y11 in a deltaic plain environment (D3) and unit Y12 in a delta (D). A large prograding trend follows this and covers units Y12 to Y15. Y12 stands for a deltaic environment, oscillating mostly between the frontdelta and the deltaic plain (D2-D3). In the next unit, Y13, deposits get sandier again and are dominated by heterolithic interbeds of siltstones and some massive thick sandstone beds. The depositional environment is a distal alluvial plain (AP1) with prograding sequences of about ten meters, which can be included in larger prograding sequences of several tens of meters. The prograding trend continues in unit Y14 that is typical of deposition in a distal to median alluvial plain environment (AP1-AP2) and in unit Y15 that was deposited in a median alluvial fan environment (AF2) with some more proximal trends. This unit is under- and overlaid by sharp shifts. It is followed by a new retrograding trend leading to the deposition of unit Y16 in a median alluvial plain environment (AP2). A new sharp shift marks the boundary to unit Y17 and shows a progradation. Units Y17 and Y19 were deposited in a proximal to median alluvial fan (AF3-AF2) and are separated by a short retrogradation to unit Y18 that was deposited in a median alluvial plain environment (AP2). Unit Y18 is characterized by layers of coalescent calcareous nodules, as well as the first calcareous impregnation of the matrix (1 - 1,5 m thick). At the top of unit Y19, coarse sandstones present a diffuse calcareous impregnation. Unit Y19 is followed by a new sharp transition to the new unit, Y20 that contains the last diffuse calcareous impregnation with silty interbeds and was deposited in a median alluvial plain environment (AP2).

Ages and complementary data of the Yaha series

According to the geological map (XBGMR, 1970) unit Y1 belongs to the late Lower Jurassic Yengisar Formation (Fig. 2.9). Units Y2 to Y6 were formed during the Middle Jurassic and units Y7 (Qigu Formation), Y8 and Y9 (Kalazha Formation) during the Upper Jurassic. Units Y10 to Y15 are considered as Lower Cretaceous deposits and units Y15 to Y20 as Upper Cretaceous deposits, while the Cretaceous-Paleocene boundary could not be determined and though the age is approximate.

On the Yaha section, measured paleocurrent directions are extremely spread and did not allow us to infer any main changes of paleoflow directions (Fig. 2.9).

Units Y1 to Y4 show evidences for a relatively wet environment, preserving organic matter in silt- and sandstones, as well as coal layers. Other pedogenic features are very rare or inexistent. The green-red boundary separating units Y6 and Y7, is marked by the first pedogenic features that could indicate a switch from wet to semi-arid conditions. From unit Y7 we observe an aridification trend. The overlaying red deposits contain many pedogenic features (root traces, pedogenic nodules, mudcracks, etc.) showing evidences for a relatively semi-arid deposition environment. They are overlaid by the calcrete and calcareous nodule-rich series, indicative of a semi-arid climate with very low sedimentation rate.

CORRELATIONS THROUGH THE TIAN SHAN AREA

1) METHODOLOGY

The sections from the northern and southern foothills of the Tian Shan Range, as well as from two intramontane basins (described in Sedimentological Analyses) show apparent equivalent sedimentologic formations (based on the geological maps, XBGMR, 1969, 1970, 1973a,b,1978a,b), mostly of Jurassic and Cretaceous age (see Geological Setting). The first step before correlating the series consists in determining the formation boundaries on the studied sections. Absolute ages in the Mesozoic series are very rare or inexistent. We attempt to use all available information from geological maps (XBGMR, 1969, 1970, 1973a,b,1978a,b) and descriptions from the literature that are based upon biostratigraphical data such as fossils, sporopollen analyses, etc. Unfortunately these data often lack accuracy and are too heterogenous to compare sediment series of different basins (see Geological Setting & Table 1). This is the reason why we determined the formation boundaries in comparing them with lithological descriptions (mostly from Sha et al., 2010 and Hendrix et al., 1992; as well as Eberth et al., 2001; Ashraf et al., 2010; Bian et al., 2010). On Figure 2.17 and 2.18, the age of the formation boundaries results from several literature reviews (Sha et al., 2010 and Hendrix et al., 1992; Ashraf et al., 2010). One should note that the extrapolation of these ages to lithological formations is questionable and these formations are certainly diachronic.

We then identify the main transgressive and regressive trends and major shifts in sediment depositional environments (see Fischer diagrams on Fig. 2.17 and 2.18) on each section. We focus on comparing global trends to rule out the influence of small local events, in order to identify the main

phases of maximum distality and proximity in depositional environment. These trends can be diachronic if they are controlled by significative tectonic activity. They may vary in space, but we suggest a first order correlation of the trend at the scale of the northern foothills, given the supposed tectonic homogeneity. Furthermore drastic changes in depositional environments (e.g. going from a proximal alluvial fan to a lacustrine environment at the Jurassic-Cretaceous boundary) can stand for tectonic activity ceasing or starting, as well as for drastic climatic change. They occur at a large scale and thus can be correlated. One should note that the sedimentation record can be discontinuous and sometimes these sharp shifts could be due to a sedimentation gap.

Disruptions in the sediment record such as laterally continuous calcrete features (see the Manas, Bayanbulak and Yaha sections) are good correlation-markers. They always occur in-between the same two equivalent formations and seem to have form at the same time. Their lateral continuity suggests that they were formed at the same time and calcretes indicate a large hiatus of sedimentation. On the contrary, unconformities (e.g. these related to the Kalazha Formation) observed in the Tian Shan area basins cannot be used to correlate these sections, because they are either discontinuous or poorly exposed.

The last step of this correlation is supported by the main aridification episodes (e.g. Eberth et al., 2001) that are synchronic at the first order (e.g. Heilbronn et al., in revision). They are observed all around the Tian Shan Range, but should be considered with caution when comparing sections on the northern and southern foothills. The Tian Shan area could have involved a significant paleo-range that could have induced different climatic conditions in the northern and southern foothills, as suggested by Hendrix et al. (1992).

2) MAIN CORRELATIONS

This correlation-work mostly concerns Jurassic series, as the underlying Triassic deposits could only be observed on the Manas section, and the overlaying Cretaceous series only on the Manas and Yaha sections. Moreover the Lower Jurassic deposits are well exposed (and accessible) only on the Manas section and in a limited way on the Yaha section. The correlation of Jurassic series is represented on Figure 2.17 and that of the younger deposits (Cretaceous and Paleogene) on Figure 2.18. The names of the formations are based on these described in the literature (see Table 1).

Lithostratigraphic constraints

The lower Middle Jurassic Xishanyao Formation consists of silt- and sandstones, with the thickest coal-bearing strata (up to few metres) and locally some calcareous mudstones. Deposits of this formation can be observed on the Wusu and Manas sections, as well as on the Yaha section, where the equivalent formation for the southern foothills is called the Kezilenuer Formation (e.g. XBGMR, 1969, 1970, 1973a,b,1978a,b; Hendrix et al., 1992; Sha et al., 2010).

The overlaying beds belong to the upper Middle Jurassic Totoun Formation and consist of grey, green or purple sand- and siltstones, locally with some conglomeratic deposits. These deposits occur on the Wusu, Manas and South Totoun sections (e.g. XBGMR, 1969, 1970, 1973a,b,1978a,b, Hendrix et al., 1992; Sha et al., 2010). Our observations lead to interpret the deposits on the Nileke section as similar to these. On the southern foothills, sediments of the Kalemake Formation on the Yaha section are also considered as equivalent to the Totoun Formation (e.g. XBGMR, 1969, 1970, 1973a,b,1978a,b ; Sha et al., 2010).

The lower Upper Jurassic Qigu Formation is usually described as red siltstones alternating with red sandstones on the four sections of the northern foothills and on the Yaha section on the southern foothills (e.g. XBGMR, 1969, 1970, 1973a,b,1978a,b, Hendrix et al., 1992; Sha et al., 2010).

The Kalazha Formation ends the Upper Jurassic period and is represented on the Wusu, Manas, South Totoun and Yaha sections by several meters to several tens of meters-thick conglomeratic deposits (e.g. XBGMR, 1969, 1970, 1973a,b,1978a,b; Hendrix et al., 1992; Eberth et al., 2001; Sha et al., 2010; Wang et al., 2013), sometimes with evidences of ephemeral sheet flow deposits. On the North Totoun section, no conglomeratic sediments can be observed but thick aeolian deposits. Those stand for deposition in an arid to semi-arid environment, similar to the conditions of deposition of the Kalazha Formation. It thus suggest that the aeolian deposits represent the distal equivalent of the alluvial fan environment of the Kalazha Formation.

Lower Cretaceous deposits are dominated by red siltstones and sandstones and directly overlay the Kalazha Formation. The formation names of the Tugulu Group differ between the northern and southern foothills, but deposits are comparable (dominated by deltaic and lacustrine environments) (e.g. XBGMR, 1969, 1970, 1973a,b,1978a,b; Hendrix et al., 1992). Upper Cretaceous deposits (Donggou Formation in the north and Qiakemakeqi Formation in the south) seem to be characterised on both side of the present day range by conglomeratic sediments and end with a several meters-thick calcrete that stands for the Cretaceous – Paleogene boundary.

One should note that there are no available data about the Wusu section in the literature. The facies exposed there are more proximal than in the other sections and thus difficult to compare with lithologies described elsewhere along the northern foothills.

It is also difficult to determine the formation boundaries in the intramontane basins because of the lack of literature on this topic. However, the most striking layer of the Bayanbulak section is a several meters-thick calcrete. We suppose that it formed at the same time then the calcrete from the Upper Cretaceous – Paleogene boundary in the northern foothills (Heilbronn et al., in revision). The calcrete is underlaid by several tens of meters thick deposits, similar to the Middle Jurassic deposits of the northern foothills. This is the reason why we suspect this calcrete to be underlaid by a fault (Jolivet et al., 2010). Our comparison of the deposits indicates that the lower units of the Bayanbulak section could be of Middle Jurassic age. The decrease of coal/organic matter-rich deposits could indicate the boundary between the equivalent Xishanyao and Totoun Formations. Red-colored units above the calcrete should be of Paleogene age.

Main depositional trends

The first global trend is prograding and covers the lower to upper Middle Jurassic on the Wusu and Manas sections (J2x à J2t), going from a lake (Wusu, unit W4) or prodelta (Manas, unit M18) to an alluvial fan (Wusu, unit W7) or median alluvial plain (Manas, unit M22). On the Wusu and Manas sections the onset of this progradation is represented by a “maximum of distality” (or maximum flooding surface - MFS) (see Figure 2.17). This trend is also recognisable on the Nileke section (going from a prodelta (unit N1), to a median alluvial plain (unit N6). It is followed by a “sharp” retrogradation to the delta or lake (on the South Totoun, Nileke and Yaha sections) or by a more progressive retrogradation to the lake (on the Manas section) or to the very distal alluvial plain (on the Wusu section).

The next global trend is again a progradation encompassing smaller retrogradations (see within the lower Upper Jurassic Qigu Formation, on the Wusu, Manas and South Totoun sections). This trend is dominated by thick alluvial or deltaic plain deposits on all sections of the northern foothills (Wusu, Manas, South and North Totoun sections). On the southern foothills, the Yaha section shows a trend dominated by a lacustrine to deltaic environment. All the sections except the two in the intramontane basins show that this trend ends with a drastic change to proximal alluvial fan deposits. The thickest deposits can be observed on the Manas section (ca. 300 m, unit M29) and laterally extend at the foothill scale (on the South Totoun section units TS7, TS8 and TS9 form a ca.

100 m-thick bar, unit WE5 on the Wusu section), as well as in the southern foothills (unit Y9 on the Yaha section). At the end of this impressive conglomeratic “wall” (also named the King Kong Wall), a very sharp transition occurs on all sections. This retrogradation goes from alluvial fan to subaqueous environments (except on the Yaha section, alluvial plain environment) (see Figure 2.18).

Cretaceous deposits can be summarized by one long prograding trend from a lake to a proximal alluvial fan environment (Manas section). On the Yaha section, the progradation in the Cretaceous deposits is very progressive going from a distal to a median alluvial plain environment, with one more distal excursion into the delta (unit Y12). On both sections (Manas and Yaha) the trend ends with the deposition of relatively thick alluvial fan conglomerates, marked by several meters-thick calcrete on the top. This calcrete is also observable on the Bayanbulak section, in the intramontane basin of the same name. Thus there are probably no Cretaceous deposits in the intramontane basins (Nileke and Bayanbulak). This is followed by a retrogradation as showed by the Tertiary series immediately overlaying the calcrete and deposited in an alluvial plain environment (Manas, Bayanbulak and Yaha sections).

As a summary, the Wusu section globally presents more proximal depositional environments than the other sections. In general, sections in the northern foothills seem to record more detailed trends than on the Yaha section in the southern foothills. In the intramontane basin, deposits from the Bayanbulak section do not show any subaqueous structures or facies, and only record deposition in rather proximal environments.

Climatic evolution

On all sections, Middle Jurassic deposits contain organic matter-rich or coal layers, indicating sediment deposition in a wet vegetated environment. In the lower Upper Jurassic many indicators show a global aridification trend occurring from a wet to semi-arid environment. Well-developed root traces and the first aeolian deposits can be observed. Except on the Wusu section, the first indicators for aridification appear immediately after the shift from subaquatic to subaerial depositional environments. This shows that the aridification probably started earlier than shown on Figure 2.17, but cannot be determined precisely.

In the upper Upper Jurassic deposits, large dune sets on the North Totoun section show evidences for an arid environment within the Junggar Basin. Cretaceous deposits seem to have formed under semi-arid conditions with standing bodies of water, such as delta and lakes. The calcrete at the end of Cretaceous time is also interpreted as deposited under semi-arid conditions (e.g. Alonso-Zarza, 2003; Heilbronn et al., in revision).

This compilation of existing data and our interpretation show that the formations of the literature based on the lithology are probably diachronic. The trends showing the evolution of the deposition environments are probably not synchronic, but can be considered as synchronic at the first order scale, given that they are probably mostly driven by the tectonic activity. When interpreting these trends, one should take into account that the sediment record may vary in the different part of the Tian Shan area (north, south, intramontane basins). Furthermore the regional aridification has been recorded in all Upper Jurassic deposits and matches well with the observed shift in the depositional environment trends. Given the uncertainties on the stratigraphy, this correlation gives a qualitative idea of the major depositional trends and offers the opportunity to be discussed in the regional tectonic frame.

PALEOGEOGRAPHIC SYNTHESIS

1) PALEOENVIRONMENTS AND TOPOGRAPHIC EVOLUTION OF THE EASTERN TIAN SHAN AREA

Figure 2.19 displays five sketch maps illustrating the Mesozoic paleoenvironmental evolution of the Chinese Tian Shan area. They are based upon the data described above (§ Sedimentology) completed with results from the literature. The aim of this work is to provide a large-scale picture of the depositional environments, topography and inferred tectonic setting at key periods during the Mesozoic.

The maps are based on the present day geometry of the main tectonic structures. The general tectonic framework of the range was established during the Permian strike-slip phase (e.g. Laurent-Charvet et al., 2002; Shu et al., 2002; Wang et al., 2007a, de Jong et al., 2009) and no major change except some shortening occurred since then (Jolivet et al., 2010). The amount of Cenozoic shortening has been estimated at 125 ± 30 km at the longitude of Manas (ca. $85,5^{\circ}$ E) and up to $203 \pm$

50 km at the longitude of Kashgar (ca. 76° E) (Fig. 2.1) (Avouac et al., 1993a). No estimate is available for the Mesozoic times and this will be discussed below.

Lower Jurassic

In the eastern Tian Shan area, Lower Jurassic sediment depositional environments go from alluvial to lacustrine, where coal-bearing strata can be formed (e.g. Hendrix et al., 1992; Shao et al., 1999; Shao et al., 2003). Sediment provenance analyses (see § 2.1.a The Manas Section; e.g. Hendrix et al., 2000; Li et al., 2004; Liu et al., 2013) indicate the dominant erosion of basement rocks during the Lower Jurassic, mostly corresponding to the erosion of the basement of the Permo-Triassic range. The early Lower Jurassic sediments consist in several tens of meter-thick conglomeratic deposits that unconformably rest on the Upper Triassic series in the South-Junggar and North-Tarim basins (e.g. Hendrix et al., 1992; this work).

In the southern margin of the Junggar Basin, some brackish water fauna have been determined and interpreted as evidence for a marine transgression from the North Tethys Ocean. They probably have been trapped in small basins that formed along normal faults (e.g. Sha et al., 2011; Pan et al., 2013). This idea requires available space to deposit these sediments close to the Tian Shan area (or paleo-range) and at the same time surrounded by flat areas where thick coal deposits can be formed. This would argue for an extensional tectonic setting. In the East-Junggar, Vincent & Allen (2001) describe the occurrence of conglomeratic deposits directly overlaying the unconformities (from Upper Triassic to Lower Cretaceous). This could be explained by the presence of normal faults, comparable to those in the North-Tarim Basin and thus local tectonic extension (Allen et al., 1991a). It is unclear when exactly during the Jurassic these faults were active in the North-Tarim Basin. The extension is confirmed in the Turfan Basin (Fig. 2.1) where it seems to control the Lower and early Middle Jurassic coal accumulation (e.g. Shao et al., 2003 and references therein), as well as in the North Tarim Basin (Korla subbasin, Fig. 2.1) where Jurassic normal faults leading to the formation of half-grabens (e.g. Allen et al., 1991a). During the Lower Jurassic the relief is relatively high in the Tian Shan area. Low temperature thermochronology results also show that the cooling and exhumation phase prevailing during the Triassic did not stop yet until the Middle Jurassic (e.g. Dumitru et al., 2001; De Grave et al., 2007; Jolivet et al., 2010).

The extensional features probably formed in a transtensional setting due to left-lateral movements on the North Tian Shan Fault, possibly caused by the suspected clockwise rotation of the Junggar block (Hou et al., 2009). This is probably itself linked to the clockwise rotation of Mongolia

(Siberia fixed) (e.g. Metelkin et al., 2004, 2007). Jurassic tectonic flower structures have been observed in the Western Junggar along the Ke-Wu Fault Belt as well as thrust faults in the Eastern Junggar along the Kelameili Fault (Fig. 2.1a) (Allen & Vincent, 1997). Another consequence of this small-scale tectonic activity driven by the rotation of the Junggar is the uplift of topographic highs in the basin, as well as the onset of the uplift of the Bogda Shan (North of the Turfan Basin, Fig. 2.1). This interpretation relies on paleocurrent direction analysis, as well as on an integrated petrographic and geochemical approach to provenance analysis (e.g. Greene et al., 2001; Shao et al., 2003 and discussion therein). The exact timing of the exhumation of the Bogda Shan is still poorly constrained. Whether the Turfan Basin was connected to the Junggar Basin in the Lower Jurassic remains controversial for some authors (e.g. Shao et al., 1999; Greene et al., 2001; Shao et al., 2003). Based upon paleocurrent directions (southward in the northern and northward in the southern foothills), Shao et al. (1999) do not image any relief in the Bogda Shan area during the Lower Jurassic. They suggest a Late Jurassic exhumation of the Bogda Shan due to the collision of the Lhasa Block. However Shao et al. (2003) observe different paleocurrent directions in the Early Jurassic (paleocurrent to the north in the northern and to the south in the southern foothills) and they interpret from the depositional environments and paleocurrent directions that the uplift of the Bogda Shan occurred before the Early Jurassic. However the Bogda Shan was separated from the Harlik Mountains (along the NE margin of the Turfan Basin) by a lake during the Lower and Middle Jurassic and in any case the scale of the Early Jurassic relief was relatively small.

The faults on the edges of the Yili Basin look like *en-echelon* structures and indicate that the basin formed in an extensional setting during the Permian that could possibly have been reactivated as normal faults during the Jurassic, given that the depositional environments seem to be similar to these described from the northern and southern foothills, at least during the Middle Jurassic (see 2.2.a) Nileke section).

Further to the South-East, Lower - Middle Jurassic left-lateral movement is also observed on the Altyn-Tagh Fault from $^{40}\text{Ar}/^{39}\text{Ar}$ and AFT thermochronology (Sobel et al., 2001). Small pull-apart basins formed along the fault. Further west, extension prevails and is expressed in the opening of the Great Caucasus Basin (e.g. Zonenshain et al., 1990).

Middle Jurassic

The Middle Jurassic Northern and Southern foothills of the range are separated by some relief (Fig. 2.19.a), as shown by paleocurrent directions to the North in the Northern foothills and to

the South in the Southern foothills (Fig. 3; e.g. Hendrix et al., 1992). The most proximal environments to the range consist in alluvial fan deposits coming from the South in the Northern foothills (on the Wusu section) and from the North in the intramontane basin (on the Bayanbulak section) (see § Sedimentology), indicating that these two sections are separated by some relief. The Bayanbulak section is situated on the northern side of the basin. No data are available to constrain the possible source from the southern margin of the Bayanbulak Basin. Lacustrine or deltaic environments with coal deposits (Fig. 2.2.6) are identified in more quiet environments located further away from the relief (North and South Totoun sections in the Northern foothills; Nileke section in the intramontane basin and Yaha section in the Southern foothills). The extent of these lakes and deltaic environments is difficult to constrain; nevertheless limestone deposits (Fig. 2.2.1) have been described in both foothills (e.g. Hendrix et al., 1992; Sobel, 1999; Jiang et al., 2008; Heilbronn et al., accepted) and stand for relatively wide and locally deep lakes. Detrital zircon provenance analyses also indicate the occurrence of some relief separating these two foothills. On the southern margin of the Junggar Basin, Yang et al. (2013) determined sources from the North Tian Shan, the Yili block, the Central Tian Shan magmatic rocks and even from the CTS block to the south. On the northern margin of the Tarim Basin, in the Kuqa sub-basin major contributions are determined from the South Tian Shan and even from the Yili-Central Tian Shan (Li & Peng, 2010; Liu et al., 2013).

In the Middle Jurassic series, Lower and Middle Jurassic volcanic zircons have been dated. We also observed Middle to Upper Jurassic tuff layers (cf. § 3; Fig. 2.2.5) on the Wusu section (and potentially on the Manas and North Totoun sections). This Mesozoic volcanic activity is unlikely to have taken place in a compressive tectonic setting, comparable to the Cretaceous-Paleogene extension and volcanism in the Tuyon Basin in NW-China and the Kyrgyz Tian Shan (Sobel et Arnaud, 2000). Little is known about the origin and sources of this volcanism (Deng et al., 2010; Wang & Gao, 2012; see discussion in: Yang et al., 2013). However, Jurassic and Cretaceous magmatism (e.g. Jolivet et al., 2007) has been reported in Central Asia (e.g. Goby Altay, Mongolia, NE-Tibet), where mantellic origin has been identified.

Furthermore the presence of Lower Jurassic synsedimentary volcanic zircons in Middle Jurassic strata shows the erosion of previously deposited sediments (Yang et al., 2013). This can be interpreted as uplift of the sediment series. Sediment recycling is also expressed in our sediment provenance analysis (see 2.1.a) Manas section; Fig. 2.3) and shows that the sediment cover has been dominantly eroded with only very little basement rocks involved. We infer that the exhumation must have been relatively low. We will see later that the sediment recycling goes on through most of the Mesozoic.

During the Middle Jurassic the rotation of the Junggar block is still active, leading to a transtensional setting prevailing in the Tian Shan area. Left-lateral movements on the major inherited faults confine the Middle Jurassic relief mainly along these zones, like during the Lower Jurassic. As a consequence of the rotation, the uplift of the Bogda Shan carries on (Fig. 2.19.a). During the Lower Jurassic, the main source rocks were of sedimentary origin, while mostly volcanic-rich lithics were eroded from the basement of the Bogda Shan in the Middle Jurassic (e.g. Greene et al., 2005). Given that the Bogda Shan basement is dominantly composed of volcanic rocks, these provenance data indicate the complete erosion of the sediment cover, followed by the initiation of the basement erosion and thus the continuous and deep exhumation of the Bogda Shan.

The Middle Jurassic marks the onset of a slow erosion period with very low AFT cooling rates (e.g. Dumitru et al., 2001; Jolivet et al., 2010). During the Middle Jurassic the topography starts to be flatter than previously, but probably not inexistent. Later this flattening will result in a general planation of the Tian Shan area (e.g. Dumitru et al., 2001; Li et al., 2004; De Grave et al., 2007; Jolivet et al., 2010). In the following, we describe some “low reliefs”. We define them as reliefs that did not generate sufficient exhumation to be thermally recorded (e.g. by AFT or (U-Th)/He analyses). In other words, we do not expect these reliefs to be higher than 1000-2000 m for a geothermal gradient of ca. 30°C/km.

In the South-Junggar and North-Tarim basins, steep tectonic structures have been imaged on seismic profiles and interpreted as inverse faults (e.g. Kuang et al., 2010; Zhang et al., 2011b). Before being reactivated and inverted, these very steep faults were possibly normal faults. Thus the Middle Jurassic tectonic evolution of the Tian Shan area could (at least locally) have been controlled by transtensive tectonic. These faults could possibly have been reactivated as inverse faults in the late Lower Cretaceous, when tectonic activity is also recorded in the widespread unconformities (in the Junggar, Turfan and Tarim Basins; Table 1; e.g. Hendrix et al., 1992). Furthermore these reactivated faults seem to be sealed by Upper Cretaceous deposits in the interpretation of Zhang et al. (2011b). The lack of good field observations does not allow us to infer if this resulted in the opening of pull-apart basins, similar to those created in Lower – Middle Jurassic along the Altyn-Tagh Fault (e.g. Sobel et al., 2001). Although pull-apart basins form in a transtensional context, they can also develop in transpressive conditions. They show that deformation is localised along the faults, while half-grabens indicate distributed extension. In this case, it would rather imply half-graben systems, similar to those described in the North-Tarim (Allen et al., 1991a).

Coal-bearing strata are widespread all over Central Asia at this period and attest of similar quiet deposition conditions (e.g. Le Heron et al., 2008 in Siberia; Sobel, 1999 in SW-Tarim; Ritts &

Biffi, 2001 in the Qaidam Basin). In Mongolia, the very low exhumation, and the good preservation of the planation surface also shows a relatively quiet setting (e.g. Vassallo et al., 2007; Jolivet et al., 2007). During the Middle Jurassic, there is no block collision occurring on the Southern Margin of Asia. The collision of the Qiangtang and Kunlun blocks occurred in the Middle Triassic (e.g. Roger et al., 2008, 2010, 2011) and most probably did not affect Central Asia anymore during the Middle Jurassic. Furthermore this quiet tectonic phase predates the collision event of the Lhasa Block in the Lower Cretaceous (e.g. Kapp et al., 2007). The Middle Jurassic general extensional tectonic setting is comparable with the one further west, where it generates the start of the opening of the South Caspian Basin (e.g. Brunet et al., 2003; Golonka, 2004).

Upper Jurassic

The Upper Jurassic deposits seem to record a large progradation phase in the sediment depositional environments (Table 2) characterised at the end of this time period by a very abrupt arrival of alluvial fan deposits around the Jurassic – Cretaceous transition (e.g. Hendrix et al., 1992; Fig. 2.17), due to a progradation peak, that suggests a renewed topography building in the Tian Shan area at this moment (Fig. 2.19.b).

Deposition mainly takes place in deltaic or alluvial plain environments and marks the shift from the Middle Jurassic quiet sedimentation phase to the late Upper Jurassic – early Lower Cretaceous alluvial fan deposition (Kalazha Formation). Coal-bearing strata disappear as well as lacustrine environments. Sedimentation also reflects the aridification trend that is recorded in the paleosol features and variations of palynoflora assemblages (e.g. Allen et al., 1991b; Hendrix et al., 1992; Eberth et al., 2001; Shao et al., 2003; Li et al., 2004; Ashraf et al., 2010). Middle to Upper Jurassic volcanic zircons still occur in the deposited sediments, as well as some very local interbedded ash deposits in the Kuqa subbasin (e.g. Li et al., 2004). They indicate an ongoing volcanic activity (e.g. Yang et al., 2013). The recycling of sedimentary strata (see Lower Jurassic) shows that some relief exists (e.g. Li et al., 2004 in the North Tarim Basin) probably in relation with normal faults.

Given that the thermal stability phase (AFT data) went on until the Lower Cretaceous, the topography must still be relatively flat.

East of Central Asia, the Upper Jurassic corresponds to a general extension phase with the onset of grabens and metamorphic core complexes, as well as the opening of several sedimentary basins in NE-China (e.g. Songliao, Hailar, Erlian, Gobi Est, Yingen et Sanjing, see Fig. 2.1) (e.g. Charles, 2010 ; Daoudene, 2011 and references therein). The most recent model proposes a double verging

subduction for the Siberia – Mongolia collision (Chapitre 1) that would induce extension in the Mongol-Okhotsk area (Daoudene, 2011). West of Central Asia, the South Caspian Basin is still opening (e.g. Brunet et al., 2003). Thus the extension phase is more widespread than during the Lower Jurassic and reached its paroxysm in the late Upper Jurassic. We suggest that the relief is slightly higher than during the Middle Jurassic, but lower than during the coming Jurassic – Cretaceous transition.

Upper Jurassic – Lower Cretaceous transition (Kalaza Formation)

The Jurassic – Cretaceous transition is marked by several meters to few hundred meters thick outstanding alluvial fan deposits in both the Northern and Southern foothills of the Tian Shan Range (see Fig. 2.17), as well as in the Turfan Basin (e.g. Hendrix et al., 1992; Wang et al., 2013). In the intramontane basins some proximal alluvial plain deposits have been observed (Fig. 2.14; 2.2.a) Nileke section and Fig. 2.14). According to the geological map (REF) they are of Lower Cretaceous age (see 2.2.a) Nileke section) and limited to a single, restricted outcrop. It is difficult to conclude on a very limited deposition, or a nearly complete erosion.

Again the paleocurrent directions on both foothills of the Tian Shan Range argue for some relief. In the Turfan Basin, the main sediment provenance area is henceforth the Bogda Shan in the North, and the Qoltag Mountains in the South become a minor source of sediments (Fig. 2.1; Shao et al., 1999; Shao et al., 2003). In the Northern foothills of the Tian Shan our in-situ sediment petrology analyses (see 2.1.a) Manas section) indicate that sediments of the Kalazha Formation are nearly exclusively the product of the erosion of previously deposited sediments. Furthermore the occurrence of Middle Jurassic magmatic zircons in the Upper Jurassic deposits (Yang et al., 2013; cf. § Middle Jurassic) shows sediment recycling. The fact that only the sediment cover has been eroded and hardly any basement rock was involved, shows that exhumation must be very low. This is in accordance with nearly null cooling rates recorded by AFT data (e.g. Dumitru et al., 2001; Jolivet et al., 2010). Nevertheless it is contradictory with the erosion products of the range that have been eroded and deposited in alluvial fans, nearly all over the area.

No volcanic series have been observed in the field at the Jurassic – Cretaceous transition, but Middle to Upper Jurassic volcanic zircons still occur in the deposits and stand for some volcanic activity (Yang et al., 2013).

Here, deposition in alluvial fans shows that the topography is generally flat and low relief is superimposed to the topography, as no exhumation has been recorded by low-temperature data (e.g. Dumitru et al., 2001; Jolivet et al., 2010). However, it remains difficult to interpret what is the reason of the brutal shift that leads to the deposition of the Kalazha formation (from alluvial plain to alluvial fan environments) (see Figures 2.17, 2.18). It is very different from a continuous and progressive progradation indicating the formation and erosion of some significative relief such as the Cenozoic progradation phase recording the reactivation of the Tian Shan Range (see in the Chinese Tian Shan the Duzhanze and Xiyu Formations in Charreau et al., 2009a; and in the Kyrgyz Tian Shan the Neogene to Quaternary deposits in Bullen et al., 2001). At the Jurassic – Cretaceous transition, unconformities seem to be local (e.g. Shao et al., 1999; Bian et al., 2010; Wang et al., 2013). If a compressive setting prevailed, such an unconformity would be a widespread striking feature in the field. Our observations do not fit in a regional flexural context, so again we expect here the presence of normal faults. In the late Upper Jurassic, the tectonic activity on normal faults could increase and thus built higher relief than previously. Their erosion could produce the thick conglomeratic deposits of the Kalazha Formation. Furthermore these extensional tectonic features fit very well to the large scale extension regime that prevails in Central Asia (e.g. Zorin, 1999; Charles et al., 2010). The Jurassic – Cretaceous transition also corresponds to a pic of extension in Siberia and Eastern Mongolia, where large Metamorphic Core Complexes developed (e.g. Donskaya et al., 2008; Daoudene et al., 2009, 2013). Furthermore extensional basins formed from the late Upper Jurassic and during the Lower Cretaceous in Mongolia (e.g. Saiki & Okubo, 2006; Erdenetsogt et al., 2009). To the West, it also matches with the maximum of extension in the South Caspian Basin and associated structures (e.g. Brunet et al., 2003; Golonka, 2004).

The intensity of the extension in the Tian Shan area remains difficult to estimate. If the extension would not increase and be comparable to the one prevailing during the Middle Jurassic, the late Upper Jurassic erosion could be driven by the intensification of the aridification. In deed we observed aeolian deposits in the Northern foothills with several meters high dunes (see South and North Totoun Sections) in the late Upper Jurassic. This is in accordance with early Lower Cretaceous aeolian deposits described from the East-Junggar area by Vincent & Allen (2001) and Eberth et al. (2001). It is difficult to imagine solely climate change as a cause of this brutal shift in deposition, given that the aridification is set up progressively since the Middle Jurassic (e.g. Eberth et al., 2001; Shao et al., 2003; Ashraf et al., 2010, Sha et al., 2011; Pan et al., 2013). Thus it is not possible to determine if the extensional setting was more influenced by the climate or the tectonic activity.

Lower Cretaceous

We infer a relatively flat topography, based on subaqueous dominated depositional environments (delta and lakes in the Northern foothills, e.g. Hou et al., 2009; Bian et al., 2010; delta and distal alluvial plain in the Southern foothills, e.g. Hendrix et al., 1992; Wang et al., 2013) with evidences of a deeper lake with anoxic lacustrine black shale further West, on the Awate River section (e.g. Hendrix et al., 1992). In the intramontane basins (Hexilagen and Bayanbulak, Fig. 2.8 & 2.7) no sediment record younger than the supposed Lower Cretaceous proximal alluvial plain sediments (see § Upper Jurassic – Lower Cretaceous transition) and older than Cenozoic deposits is available. In the Flaming Mountains of the Turfan Basin, the observations of Hendrix et al. (1992) differ from those of Shao et al. (1999). The first authors do not observe any lake deposit, whereas Shao et al. (1999) interpret deposition in deltaic lake and lake environments. We can suppose that the exact locations of their outcrops are separated by lateral variation of facies and depositional environments and that the lakes might be of small extension. Thus no evidence for tectonic activity has been recorded.

One should note that the Kalazha Formation or equivalent thick-bedded alluvial fan deposits occurring close to the Upper Jurassic – Lower Cretaceous transition in the Tian Shan area (also in the NE-Junggar and Turfan basins) lacks of biomarkers and has been attributed to different ages: either a late Upper Jurassic age (e.g. Hendrix et al., 1992, 2000; Eberth et al., 2001; Jin et al., 2008) or a Lower Cretaceous age (e.g. Vincent & Allen, 2001; Li et al., 2004; Wang et al., 2013). This problem will remain difficult to solve, since the Kalazha is a diachronic formation, that probably contains internal unconformities (e.g. Eberth et al., 2001; Vincent et Allen, 2001).

Even if continuous cooling of the basement occurs during the Cretaceous as recorded further West in the Kyrgyz Tian Shan (e.g. De Grave et al., 2007; De Grave et al., 2013), no or only poor record of a Cretaceous exhumation phase is reported in the Chinese Tian Shan (Dumitru et al., 2001; Jolivet et al., 2010). This very little exhumation is coherent with the preservation of the Jurassic planation surface in the range as farther in Central Asia (in Mongolia: Vassallo et al., 2007; Jolivet et al., 2007; in Siberia: De Grave & Van den haute, 2007; Jolivet et al., 2009, 2011, 2013a), as well as in North-Tibet (Jolivet et al., 2001 ; Roger et al., 2011). The continuous occurrence of Jurassic and Lower Cretaceous volcanic zircons in the Lower Cretaceous deposits again indicates recycling of the previously deposited series (e.g. Yang et al., 2013).

There are neither evidence of compression nor of extension during this period. Solely the general geodynamic context shows wide-spread extension. It seems that the Upper Jurassic extension stopped abruptly after the Jurassic – Cretaceous transition. We thus suggest that the Lower Cretaceous period is marked by an intermediate geodynamic phase following general and widespread extension, and preceding a new reactivation phase of the range, as shown by the continuous cooling recorded throughout the Cretaceous (AFT data, e.g. De Grave et al., 2013). Central Asia is separated from the South Asia margin by several blocks and thus may have recorded the Lhasa block collision with some delay. We suggest that the end of extension in Central Asia is possibly caused by the collision of the Lhasa block on the Southern margin of Asia during the early Lower Cretaceous (e.g. Kapp et al., 2007). Given that the lag time between the onset of the collision (ca. 140-130 Ma) and the end of the tectonic activity in the Tian Shan area (extension) is probably relatively short, we can imagine that the end of the deposition of the Kalazha Formation could be synchronic through the Tian Shan area and mark the end of widespread extension, as well as the start of the collision of the Lhasa Block.

Thus this interpretation shows that the deposition of the Kalazha Formation cannot be explained by the reactivation of the range due to the Lhasa collision far-field effects as accepted until now (e.g. Hendrix et al., 1992).

We propose that the Kalazha Formation is the result of the erosion of less than 2-3 km high relief (not recorded by AFT data) formed in relation to normal faults in a general extensional tectonic setting. Indeed if extension ceases, there is no more associated relief that can be eroded. This leads to a drastic drop in sediment flux, while the accommodation space created during the Jurassic still exists. During the Lower Cretaceous this space is filled with water and shows the lack of (or very poor) sediment supply and should explain the occurrence of large lakes in the basins, at least in the South-Junggar Basin. In the field there is no evidence for any climate change at the Jurassic – Cretaceous boundary or during the Lower Cretaceous that could explain the formation of such large water-bodies.

Upper Cretaceous

In the South-Junggar, North-Tarim and Turfan basins, Upper Cretaceous series consists mostly of alluvial fan deposits unconformably overlaying the Lower Cretaceous series (Fig. 2.3 and 2.9, e.g. Hendrix et al., 1992). Though our observations on the Yaha section shows that the Upper Cretaceous deposits start and end with alluvial fan deposits, they are dominantly composed of

alluvial plain deposits. Li et al. (2004) mentioned that it is generally accepted that Upper Cretaceous deposits do not exist in the Kuqa sub-basin. Based on stratigraphic correlations with the Junggar Basin, this contradicts our observations (Fig. 2.9) and the ones in Hendrix et al. (1992). The first-order progradational trends are similar on the Manas and Yaha sections, and we infer that Upper Cretaceous sediments were effectively deposited in the Kuqa sub-basin (Fig. 2.18).

The sharp shift from alluvial plain to alluvial fan occurring at the Lower – Upper Cretaceous transition on the Manas section (northern foothills) and on the Yaha section (southern foothills) is similar to the one occurring in the late Upper Jurassic (base of the Kalazha Formation) and suggests a renewed topography building in the Tian Shan area at this moment, as well as tectonic activity that is attested by the widespread unconformity observed in the basins (Junggar Basin, Hendrix et al., 1992; Eberth et al., 2001; Turfan Basin, Hendrix et al., 1992; Tarim Basin, Hendrix et al., 1992; Wang et al., 2013). This would not argue for an extensional setting, but rather for compression.

Upper Cretaceous deposition in alluvial plain environments on the Yaha section seems to show that the section was located in a more distal part to the range than the Manas section in the northern foothills (Fig. 2.18 and 2.19.C). Similarly, on the North Totoun section the sediment deposition during the Upper Cretaceous seem to be restricted to lake environments throughout the Cretaceous (Fig. 2.5). This shows that a lake environment occurs relatively close to the range and that the relief is mostly located along the major fault zones in the central part of the present-day range (e.g. North Tian Shan Fault, Nikolaev Line and Narat Fault, South Tian Shan Fault) as shown by exhumation ages around 65-60 Ma on the Narat Fault for instance (e.g. Wang et al., 2008; Jolivet et al., 2010) (Fig. 2.1).

In alluvial fan deposits, paleocurrent directions are not easy to determine, though they seem to go from south to north in the northern foothills and the other way around in the southern foothills (Fig. 2.9; e.g. Hendrix et al., 1992), indicating erosion of the relief in-between. There is no evidence for deposition inside of the range during the Upper Cretaceous (Hexilagen and Bayanbulak Basins) (XBGMR, 1969, 1973b). Furthermore sediment petrology analyses (Fig. 2.3) indicate that the proportion of basement rocks increases again during the Upper Cretaceous, showing that the sediment cover has been stripped off and sediment recycling is not possible anymore. Together with the cooling phase (e.g. AFT ages around 100 Ma along the Narat Fault in the Central Chinese Tian Shan, Dumitru et al., 2001; Jolivet et al., 2010) that started in the late Lower Cretaceous and goes on during the Upper Cretaceous, this shows an Upper Cretaceous reactivation episode of the range.

The reactivation goes along with a tranpressive tectonic setting, with left-lateral movements on the strike-slip faults (Jolivet et al., 2010). It probably induces small scale thrust faults along the relief and confirms the presence of a significant relief in the Tian Shan area during the Upper Cretaceous.

The very abrupt transition from the alluvial fan to the lake deposits at the Upper Jurassic – Lower Cretaceous transition is noteworthy (Manas, South Totoun and Yaha sections; Fig. 2.17 & 2.18). This could express a sharp change either in the tectonic setting or in the climatic conditions. Far field-effects of the collision of two blocks are often delayed, such as the Cenozoic reactivation of the Tian Shan Range (e.g. Avouac et al., 1993a; Charreau et al., 2009a; Bullen et al., 2001) that is recorded 30 - 40 Ma after the beginning of the collision between India and Asia. In the same way, we suggest that the Lower – Upper Cretaceous transition and reactivation of the Tian Shan Range results of the accretion event of the Lhasa block 20 to 40 Ma years earlier. This does not seem to affect the clockwise rotation of the Junggar block that reaches its paroxysm during the Lower Cretaceous (mentioned by Hou et al., 2009), due to movements of Mongolia compared to stable Siberia.

Further south-west, in the West-Tarim Basin, Upper Cretaceous deposits are marine (e.g. Sobel, 1999). This shows that the base level of the rivers was situated only a few hundreds of kilometres away. This leads to the conclusion that high reliefs could not occur in the area, at least not associated to quiet depositional environments.

Locally AFT data record exhumation around the Cretaceous – Tertiary boundary, and shows local uplift, such as the onset of the uplift of the Bogda Shan around 65 Ma (Wang et al., 2008). This could be linked to far-field effects of terrane accretions to the southwest (Kohistan-Dras Arc, Kabul Block) or to the final closure of the Mongol-Okhotsk Ocean to the northeast (e.g. Searle, 1991; Hendrix et al., 1992; Wang et al., 2008; Jolivet et al., 2010). Although the ages of the sedimentological formations are poorly constrained, the late Upper Cretaceous alluvial fan deposits observed on the Yaha section seem to parallel the reactivation phase we observe in the AFT data around 65-60 Ma (e.g. Wang et al., 2008; Jolivet et al., 2010).

Late Upper Cretaceous – Early Paleogene

On the Manas section in the Northern foothills and on the Yaha section in the Southern foothills, the Cretaceous – Paleogene boundary is characterised by a several meters-thick calcrete layer (see Chap. 4; Heilbronn et al., in review). On the North Totoun section, there is no evidence of alluvial fan deposits throughout the whole Cretaceous but solely lacustrine series, according to the

geological map (XBGMR, 1978b) and our field observations (Fig. 2.5). This could simply be related to the more distal location of this section in relation to the paleo-relief. The large lake (with wave ripples, Fig. 2.5) lasted in spite of the semi-arid to arid climate (paleosols on the Manas and Yaha sections). Hence it is very difficult to record any sedimentation hiatus in these lacustrine sediments, even less an equivalent of a calcrete. A calcrete layer has been observed on the Bayanbulak section in the intramontane basin and could have possibly formed at the same time (Heilbronn et al., in review). Vincent & Allen (2001) observed extensive paleosol development in the East-Junggar Basin at the same period. However, no report exists of such a pedogenic feature in the Turfan Basin, east of the Tian Shan Range. In the West Tian Shan, Fortuna (1983) describes the Chonkurchak Formation in the Issyk Kul Basin (Kyrgystan), a few tens of meters-thick condensed formation, deposited from the Late Cretaceous to Early Oligocene and containing paleosols and lacustrine deposits.

These paleosol features indicate environments without any sedimentation during its formation. We thus infer that the topography must have been nearly flat and the relief very low during late Upper Cretaceous-Paleogene time. This corresponds to a tectonically quiescent period extending from late Upper Cretaceous through much of the early Cenozoic with very low cooling rates (De Grave et al., 2013; Macaulay et al., in review, see Chapter 3 in Macaulay, 2013). This phase certainly lacks of temporal constraints, but shows a very quiet time laps ranging from late Upper Cretaceous to Paleogene, before the major Cenozoic event occurring on the South Asia margin, the India-Asia collision.

Given that AFT data locally record exhumation around 65-60 Ma (e.g. Bogda Shan, Wang et al., 2008) followed by a tectonic quiescent phase, given that calcrete has been observed on the Manas section not so far from the Bogda Shan and that the calcrete is a synchronic feature over the range, it appears that the calcrete potentially formed during the Paleocene. The calcrete has not been observed in the Turfan Basin, possibly because the Bogda Shan was being eroded and the sediment supply prevented the formation of such a feature.

5.2.) PROGRESSIVE MESOZOIC ARIDIFICATION IN THE TIAN SHAN AREA

The climate changed progressively from humid (during Late Triassic and Early Jurassic) to the onset of dry seasons at the end of the Lower Jurassic. The clear change to a semi-arid - arid climate occurs during the Late Jurassic and can be observed all over north-west China (Allen et al., 1991b; Hendrix et al., 1992; Parrish et al., 1993; Shao et al., 2003; Li et al., 2004; Eberth et al., 2011; Pan et al., 2013). Our observations confirm this aridification trend. Thus the initiation of aridification is

difficult to determine, since it is only recorded (e.g. by paleosol features) in subaerial deposits and systematically preceded by deltaic or lacustrine sediments that record this kind of change with more difficulty.

However, the widespread semi-arid to arid climate is confirmed by our observations of Late Jurassic – Early Cretaceous aeolian deposits in the northern foothills of the Tian Shan Range (Fig. 2.4 & 2.5), similar to the observation of Eberth et al. (2001) and Vincent & Allen (2001) in the eastern part of the Junggar Basin. The large dunes indicate a peak of aridity compared to previous and following periods, apart from nowadays. We suggest that these deposits are synchronous with the alluvial fan deposits observed on the sections nearby (e.g. Manas section, South Totoun section), further in the southern foothills (Yaha section) and in the Turfan Basin (e.g. Hendrix et al., 1992). It would be tempting to associate the deposition of the thick Kalazha Formation to a climate change. Though a sole climate change cannot generate a drastic change of granulometry, such as the one observable at the base of the Kalazha Formation, as well as an angular unconformity (late Upper Jurassic) (e.g. Eberth et al., 2001).

Hendrix et al. (1992) suggested a paleo-rain shadow during the Early Cretaceous that would imply a higher evaporation and dryer conditions in the North-Tarim Basin compared to the South-Junggar basin. Our inferences from combining sedimentology and low temperature thermochronology results show that the topography was rather flat and the relief low during that period (see also our results from geochemistry data Heilbronn et al., in review – Chap. 4). Palynological analyses from the Western Tarim Basin show an arid climate since the Late Cretaceous (Wang et al., 1990a,b; Zhang and Zhan, 1991).

CONCLUSION

The aim of this study is to reconstruct the Mesozoic evolution of the paleotopography in the Tian Shan area. In this paper we combine two different approaches to solve this apparent paradox: the record of the erosion of the range in the basin (sedimentology) and the record of the exhumation of the range (low temperature thermochronology). We compared the deposits along the northern and southern foothills, as well as from intramontane basins (Yili, Bayanbulak) in order to better constrain the extent of the paleo-relief.

During the Lower Jurassic, the Upper Paleozoic relief is eroded as shown by the exhumation phase recorded by low-temperature thermochronology analyses. The Middle Jurassic period records relatively quiet sediment deposition and nearly no exhumation. The topography is relatively flat.

However, volcanic activity and sediment recycling initiates in that time and indicate the formation of small scale relief, due to normal faults. This period follows the collision of the Qiangtang block further south and fits into a general extensive setting, where the South Caspian Basin starts its opening further west.

During the late Upper Jurassic, thick alluvial fan deposits occur and indicate the denudation of newly created topography in the area. Nevertheless, the exhumation in the range is not sufficient to be recorded by AFT analyses, though these local reliefs are located along major faults and were relatively low (<2-3 km). We suggest that they occurred in association with tectonic activity on normal faults that is coherent with the regional setting, where extension is widespread all over Central Asia, from the Mongol-Okhotsk suture zone to the Caspian area.

From Lower Cretaceous, quiet deposition environments occur associated to a flat topography. Widespread extension seems to have stopped. This could be linked to the collision of the Lhasa Block (140-120 Ma) on the southern margin of Asia. During the Upper Cretaceous, relief erosion is recorded by local alluvial fans in the proximal area to the range. Continuous cooling (100-65 Ma) indicates some reactivation of the range and the presence of a significant relief. This was induced in a transpressive tectonic setting, possibly as a delayed far-field effect of the Lhasa collision.

The acceleration of exhumation (65-60 Ma) that shows a stronger reactivation phase at the late Upper Cretaceous – Paleogene transition, is seen as a response to the terranes accretions (Kohistan-Dras Arc or Kabul Block). This is followed by calcrete formation, tectonic calm and flat topography in the early Paleocene.

ACKNOWLEDGMENTS

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Cf. References at the end of the thesis

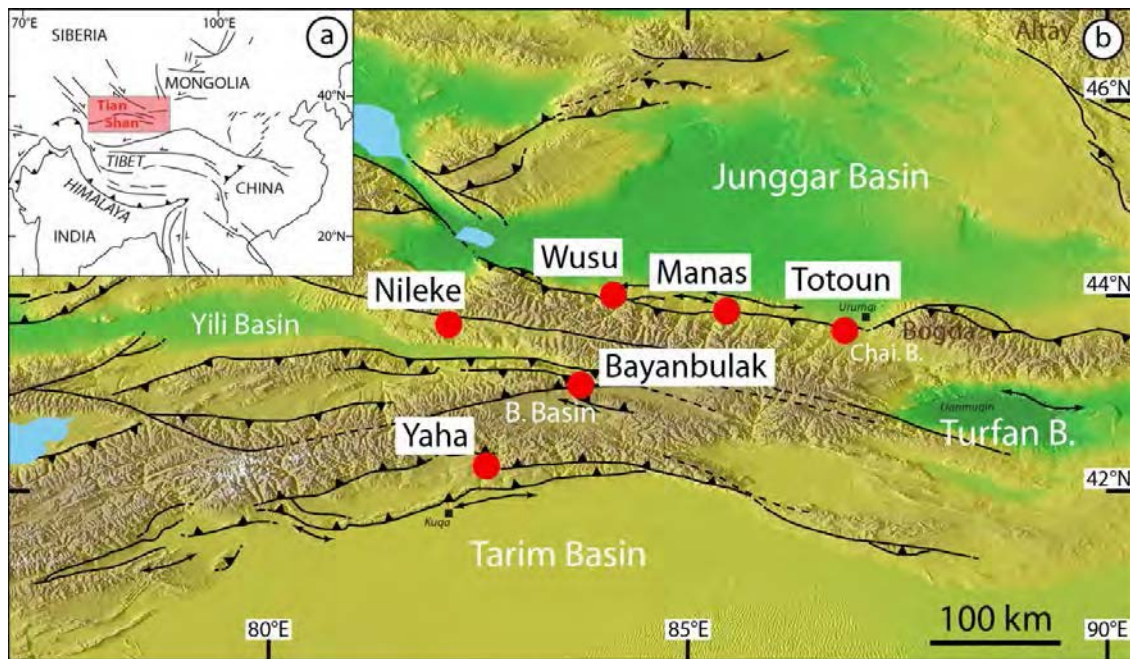


Figure 2.1 : General map of Asia to situate the Tian Shan area (a). Location of the studied sections in the Tian Shan range (b).

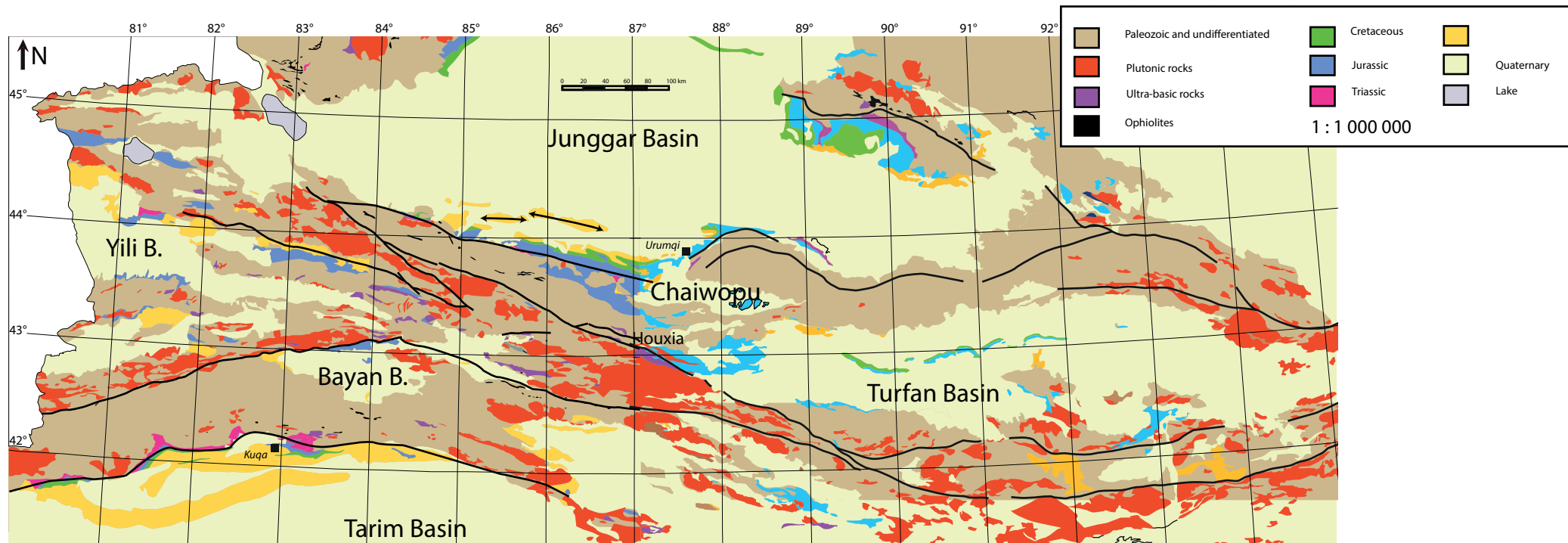


Figure 2.1.c : Geological map of the Chinese Tian Shan, redrawn after Xian et al., 2007).

Fig. 2.2.1

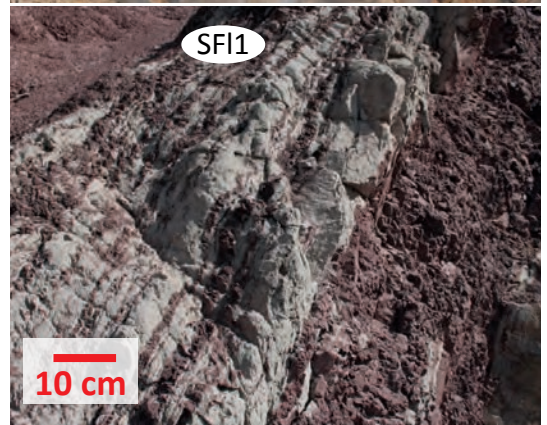
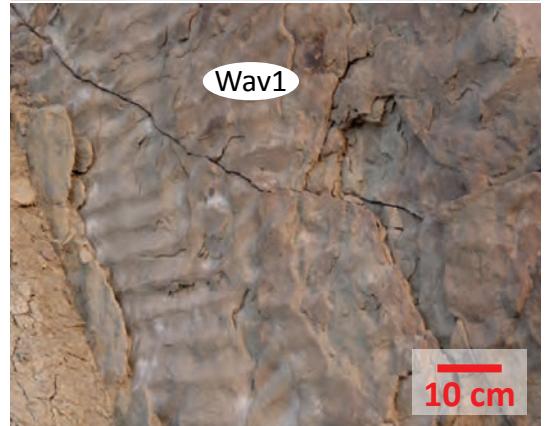
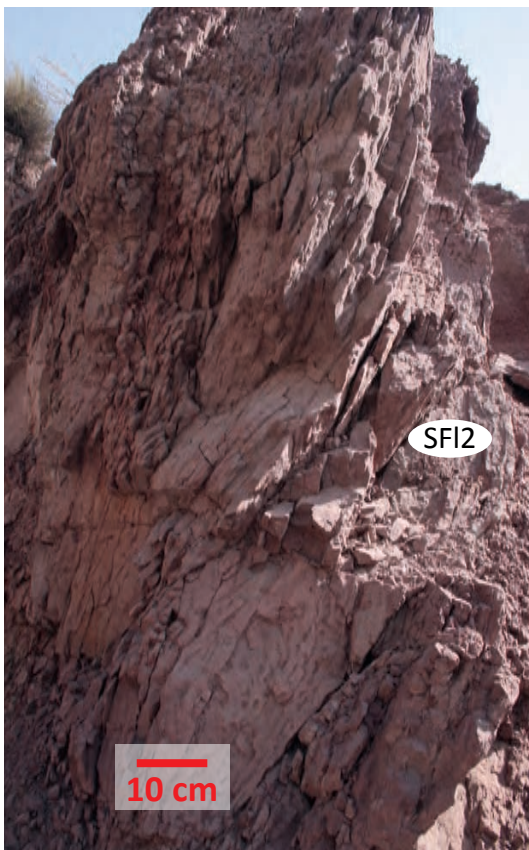
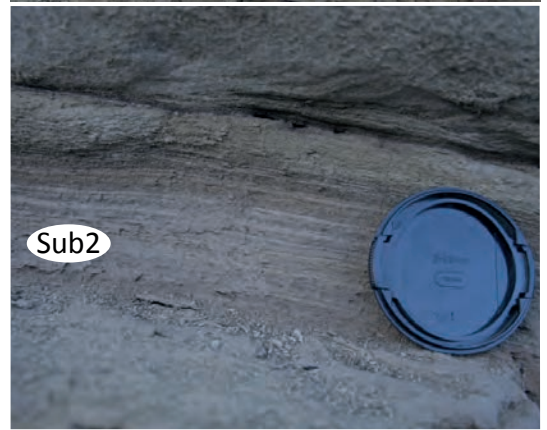
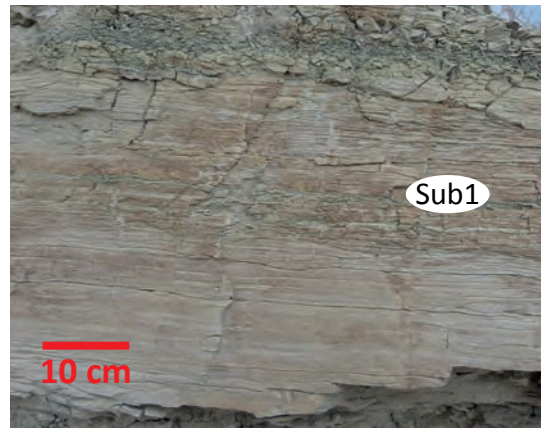


Fig. 2.2.2

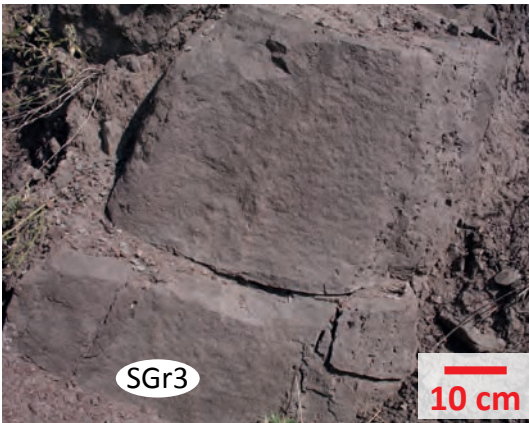
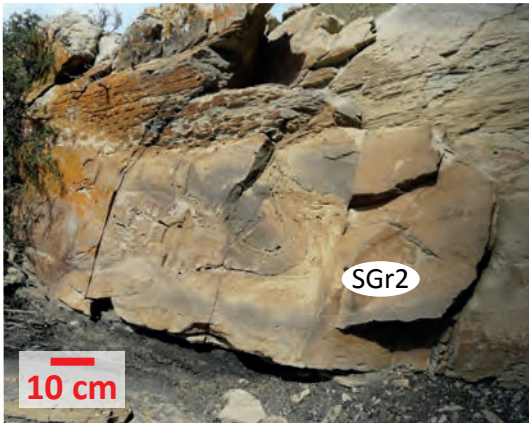
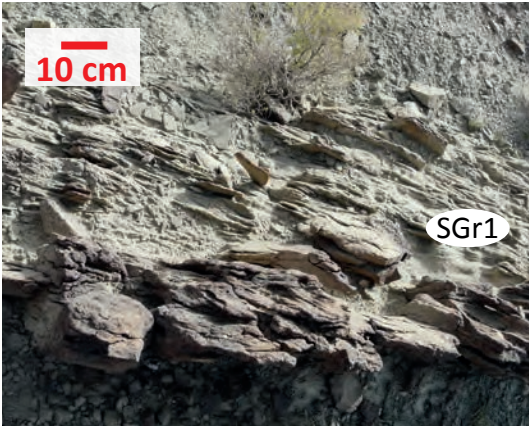


Fig. 2.2.3

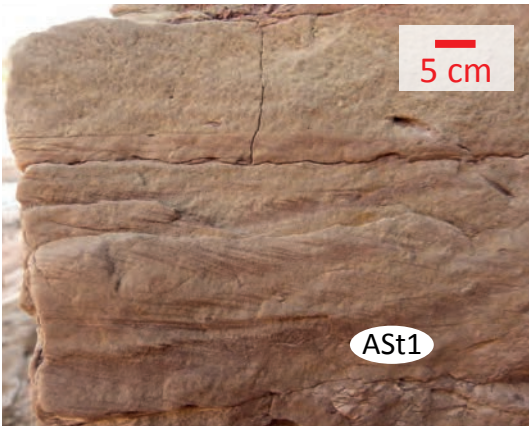
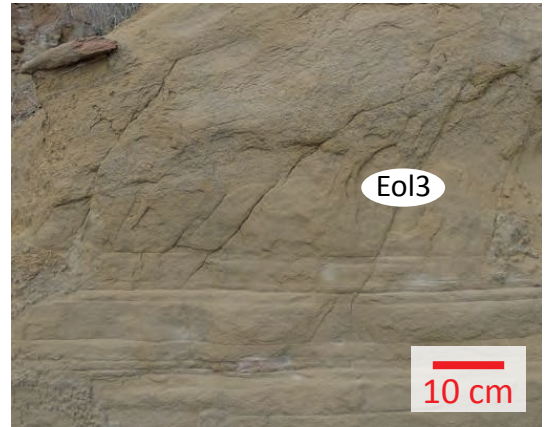
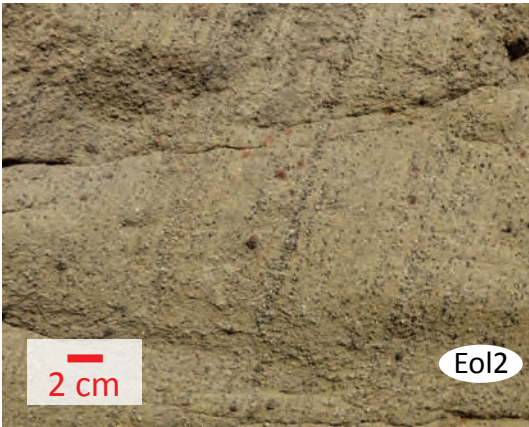
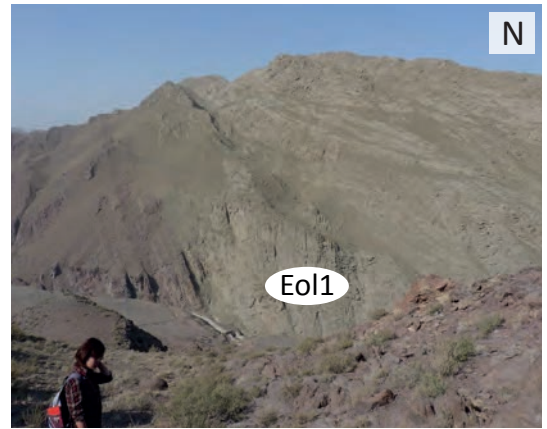
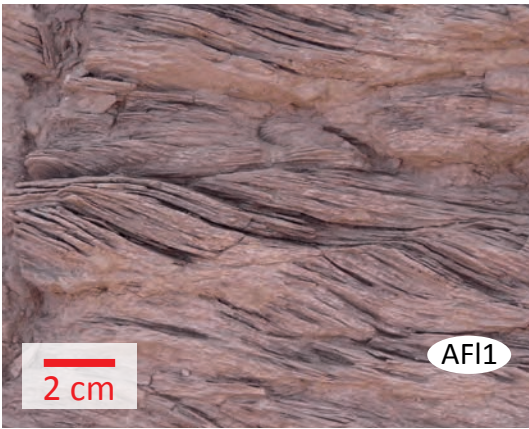


Fig. 2.2.4

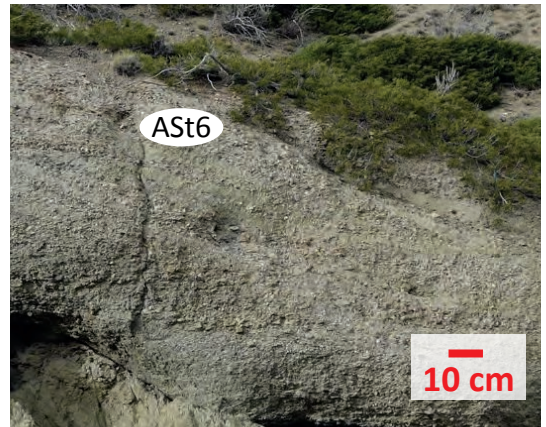
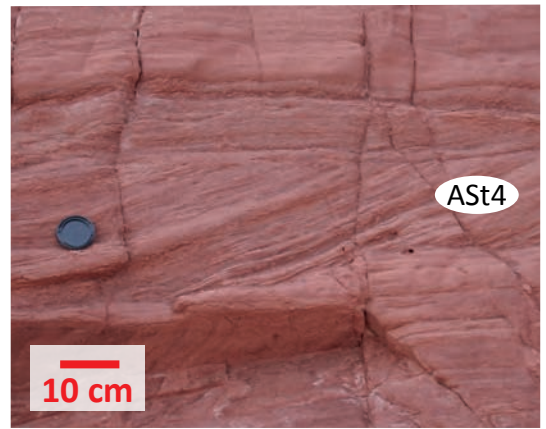
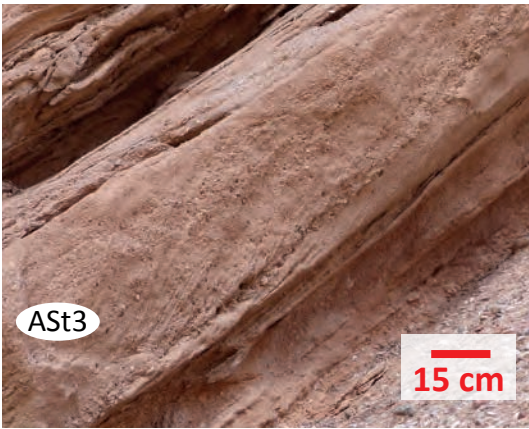


Fig. 2.2.5

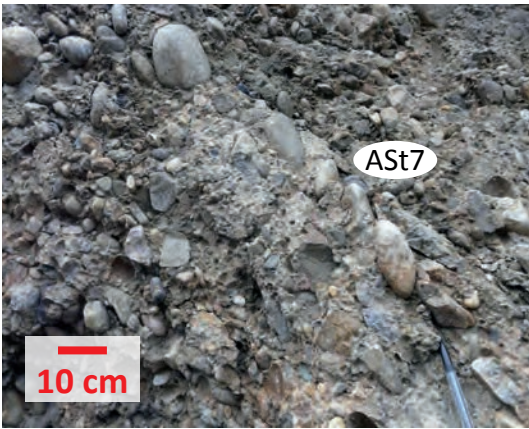


Fig. 2.2.6: Other facies

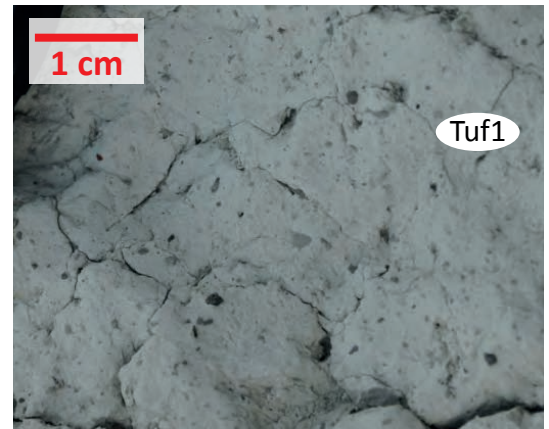
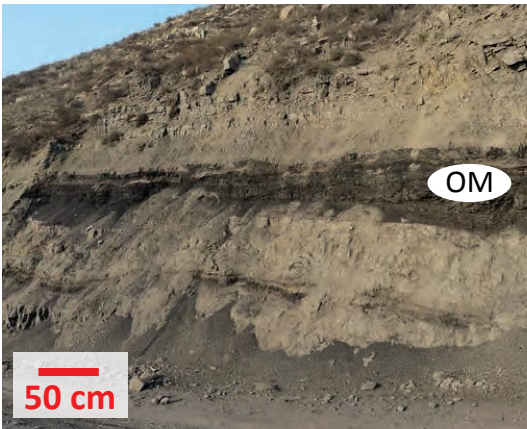


Figure 2.2.1 Most common facies occurring in the lake and delta environments described in this study (for further details see descriptions in Table 3).

Cal1: Homolithic cm to several dm-thick silty carbonate to calcareous siltstone or sandstone.

Sub1: Homolithic, several cm to several m-thick muddy to sandy siltstone, with massive to horizontal laminations.

Sub2: Homolithic, several cm to several dm-thick muddy to sandy siltstone, locally organic matter-rich, with massive to horizontal laminations.

Wav1: Cm to dm-thick fine to medium sandstone with symmetrical ripples.

SFl1: Cm to several m-thick heterolithics composed of cm to several m-thick siltstone and cm to several dm-thick fine sandstone; massive or with horizontal laminations.

SFl2: Heterolithic, several dm to several m-thick fine to coarse sandstone with mega-trough cross-bedding or horizontal planar bedding, with local strong bioturbation within the cross-bedding.

Figure 2.2.2 Most common facies occurring in the lake and delta environments described in this study (continued).

SGr1: Heterolithic, cm to several dm-thick fine sandstone to granule conglomerate, with normal grading and asymmetrical ripples to mega-ripples.

SGr2: Homolithic, dm to m-thick siltstone to coarse sandstone with soft-sediment deformation.

SGr3: Homolithic, several dm to several m-thick fine to coarse sandstone, with dewatering structures, mud rip-up clasts, local bioturbation and flute casts on the base.

AGr1: Meter to several m-thick, poorly sorted clast-supported conglomerates going gradually into matrix-supported conglomerates with a sandy matrix.

Figure 2.2.3 Most common facies occurring in the alluvial environments described in this study (for further details see descriptions in Table 3).

AF1: Homolithic, several cm to several dm-thick fine to coarse sandstone with climbing ripples.

Eol1: Several dm to several m-thick, well to moderately sorted, fine to coarse sandstone with sandflow and grainfall laminations and inverse climbing ripples.

Eol2: Well to moderately sorted, fine to coarse sandstone with planar lamination or very-low angle cm-scale cross-laminations and inverse climbing ripples.

Eol3: Well sorted, fine to medium sandstone with fine crude or irregular cm-horizontal lamination, adhesion plane beds, warts and ripples, and inverse climbing ripples.

ASt1: Dm to several dm-thick homolithic fine to coarse sandstones with amalgamated cm to several cm-thick asymmetrical ripples.

ASt2: Several cm to several dm-thick heterolithic of siltstones to fine sandstone and fine to coarse sandstone, with planar laminations or asymmetrical ripples, and locally interbedded with organic matter-rich layers.

Figure 2.2.4 Most common facies occurring in the alluvial environments described in this study (continued).

ASt3: Several dm to several m-thick fine to coarse sandstone and small pebbles, in continuous or m to several m-wide lenticular beds, with trough cross-bedding to mega-ripples and erosional basal boundaries. The foresets are often underlined by granule to pebble lags.

ASt4: Homolithic, several dm to several m-thick, fine to medium sandstone in continuous layers or locally in dm to several m-wide lenticular beds, occurring in a succession of stacked planar and trough cross-bedding.

ASt5: Homolithic, several dm to several m-thick fine to coarse sandstone with planar laminations

ASt6: Clast-supported conglomerates composed of subangular to subrounded pebbles, locally with coal or silt- to sandstone lenses. It occurs in several dm to m-thick continuous layers or several m-wide lenticular beds and can be massive or with mega-ripples, and locally contains winnowed surfaces.

Figure 2.2.5 Most common facies occurring in the alluvial environments described in this study (continued).

ASt7: Clast-supported conglomerates composed of subangular to subrounded pebbles to cobbles, with faint planar bedding or faint low-angle cross-bedding, and with pebble imbrications.

ASt8: Clast-supported conglomerates composed of angular to subrounded pebbles to cobbles, locally graded, with faint planar bedding or faint low-angle cross-bedding, pebble imbrications and locally winnowed surfaces.

AGr2: Massive, poorly sorted, matrix-supported conglomerates composed of subangular to subrounded granules to cobbles.

Figure 2.2.6 Most common facies occurring in the alluvial environments described in this study (continued).

OM: Organic matter-rich mud- and siltstone or sandstone, locally cm to dm-thick continuous coal layers or lenticular beds, with plant remains and root casts.

Tuf1: Massive, several cm to several dm-thick laterally continuous layers of fine white sandstone with small floating granules and locally plant remains, resulting of subaerial volcanic ash deposition.

Ped1: Cm to several dm-long root casts, perpendicular to the bedding.

Ped2: Loose calcareous nodules (also called “caliches”) to coalescent beds.

Ped3: Massive, several m-thick pebble beds, with coarse sandstone-matrix and pervasive calcareous impregnation/cementation.

KEY

Supposed climatic environment







 non-defined	 Wet, but not necessary standing water
 Standing body of water	 Wet
	 Semi-arid


























 FELSIC rocks

 MAFIC rocks

 TERRIGENEOUS

 LIMESTONE

Conglomerate	
Silt- and sandstone	
Shale	
Coal	
Limestone	
Calcrete	

	Horizontal planar lamination		Calcareous nodule
	Megaripples (Trough cross-bedding)		Vertical burrow
	Ripples (Trough cross-bedding)		Horizontal burrow
	Climbing ripples		Soft-sediment deformation
	Symmetrical ripples		Dewatering structures
	Plants fragments		Clay chips
	Root traces (rare to very frequent)		Normal grading
	Bioturbation (poor to strong)		Lense or channel
	Shell remains		Amalgamated lenses or channels
	Mudcracks		Coal chips
	Amalgamated megaripples		Calcareous bar
	Erosive basal boundary		
	Coalescent calcareous nodules		
	Fault		

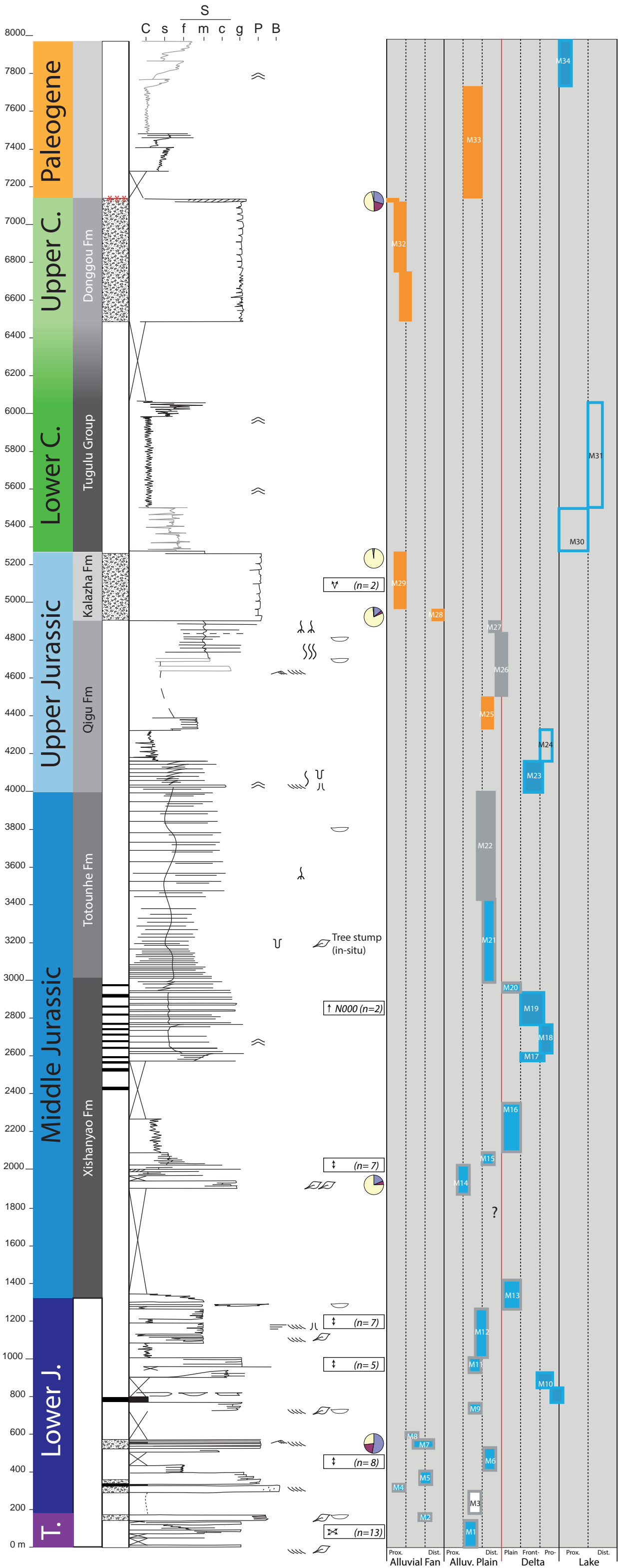


Figure 2.3: Log of the Mesozoic succession in the area of the Ningjia and Manas rivers, northern Tian Shan foothills, with paleocurrent directions, depositional environment interpretations (from alluvial fan to lake) and supposed climatic context. Scale 1/20 000.

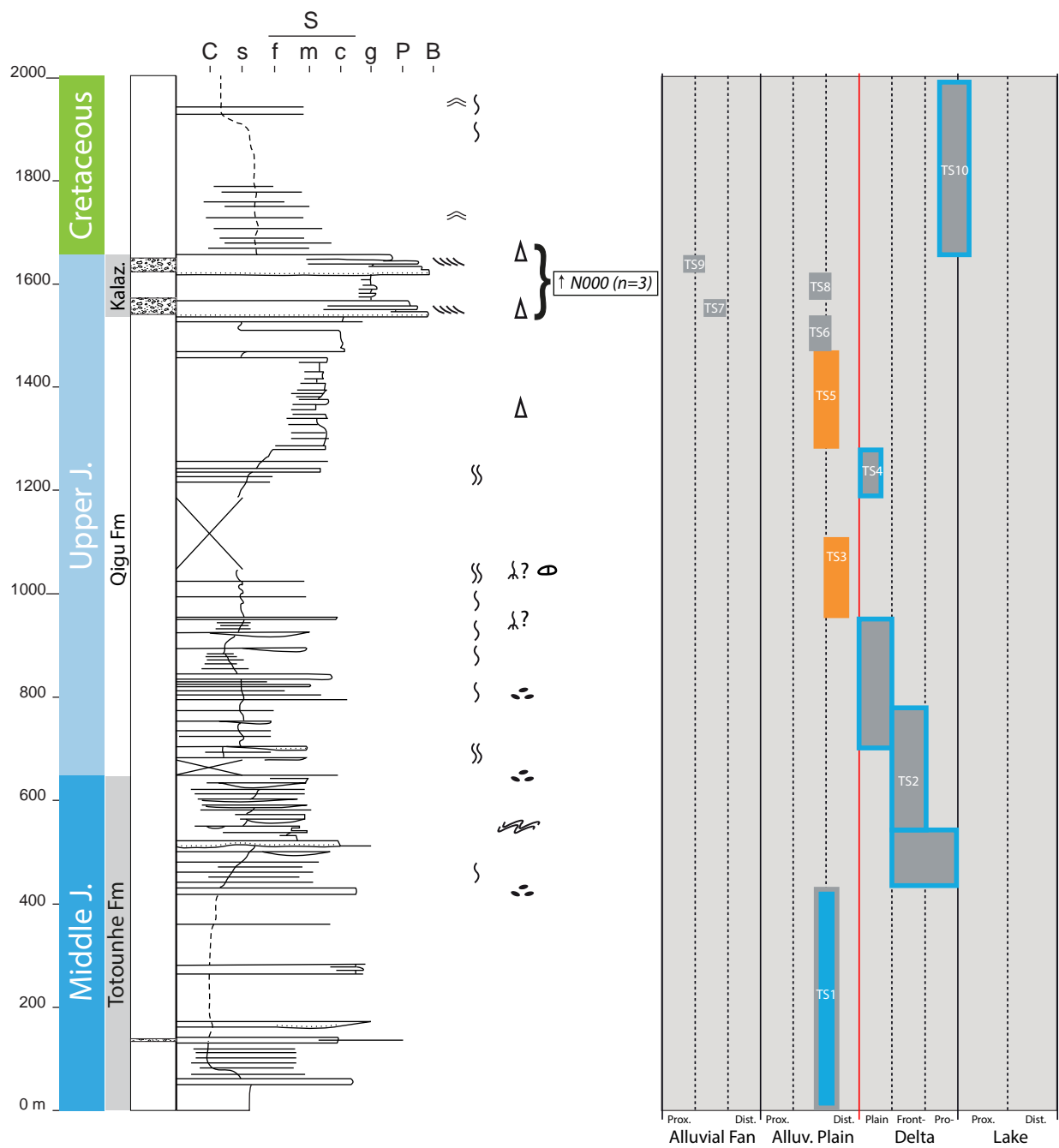


Figure 2.4: Log of the Mesozoic succession in the southern area of the Totoun river, northern Tian Shan foothills, with paleocurrent directions, depositional environment interpretations (from alluvial fan to lake) and supposed climatic context. Scale 1/10 000.

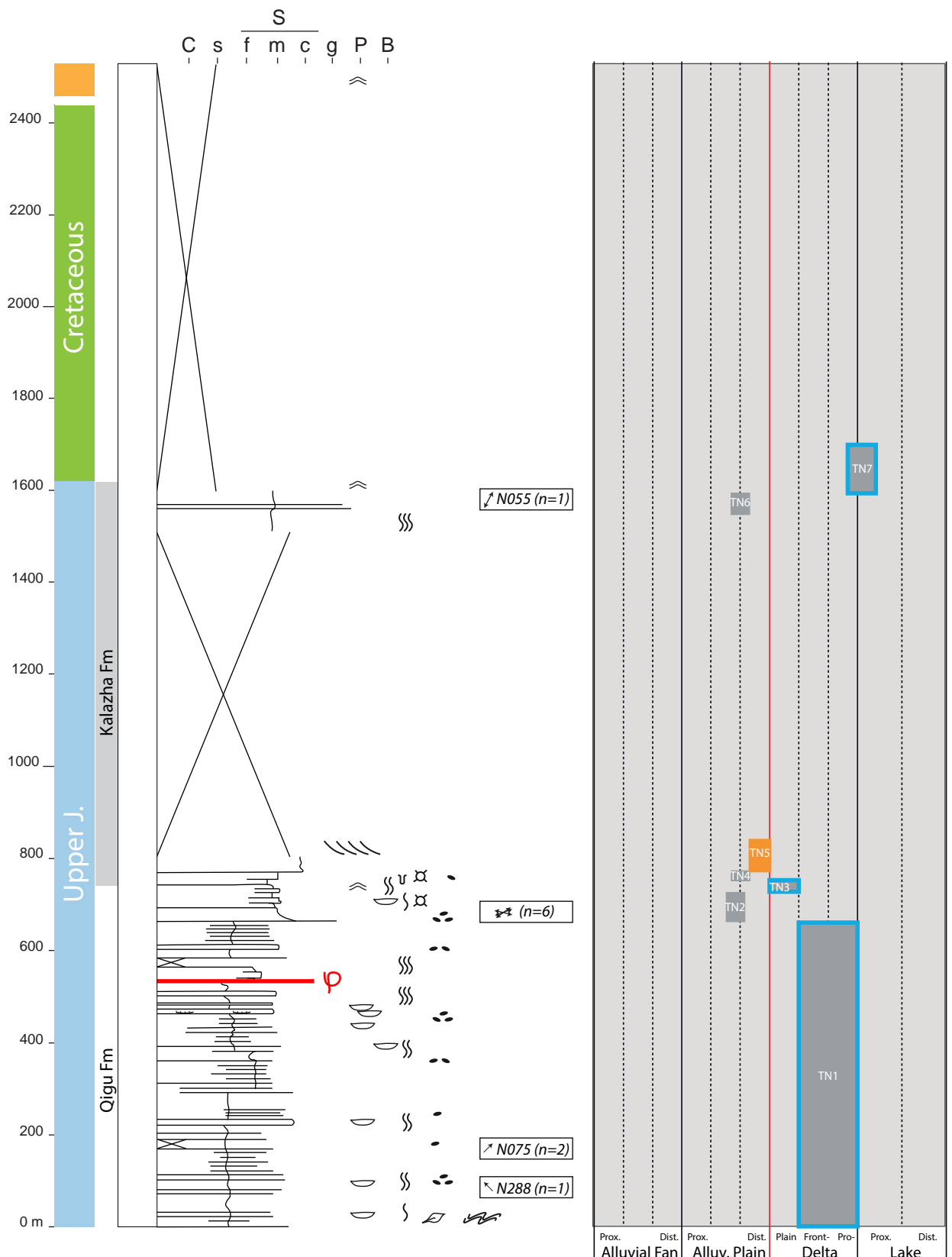


Figure 2.5: Log of the Mesozoic succession in the northern area of the Totoun river, northern Tian Shan foothills, with paleocurrent directions, depositional environment interpretations (from alluvial plain to lake) and supposed climatic context. Scale 1/10 000.

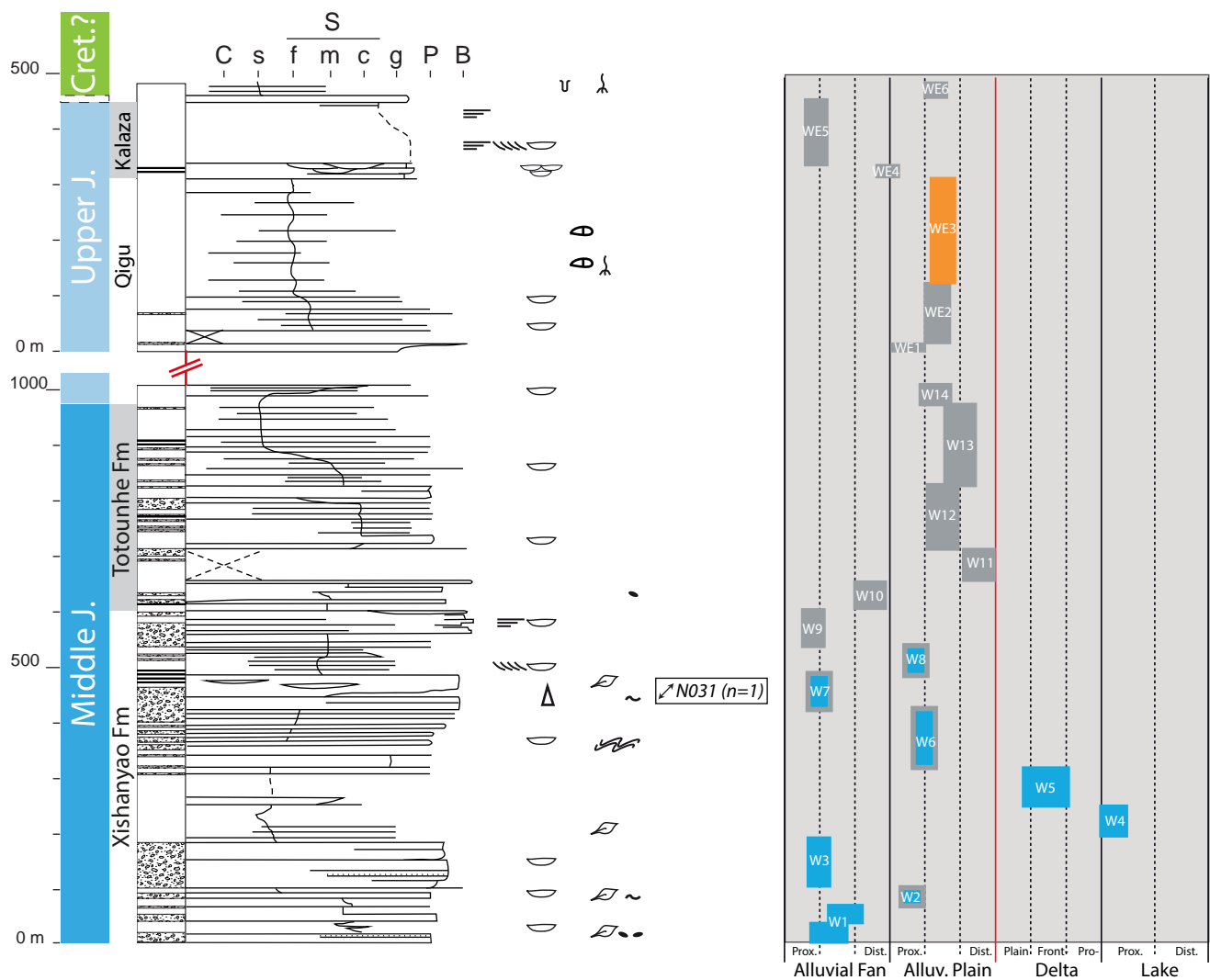


Figure 2.6: Log of the Mesozoic succession nearby the city of Wusu, northern Tian Shan foothills, with paleocurrent directions, depositional environment interpretations (from alluvial fan to lake) and supposed climatic context. Scale 1/10 000.

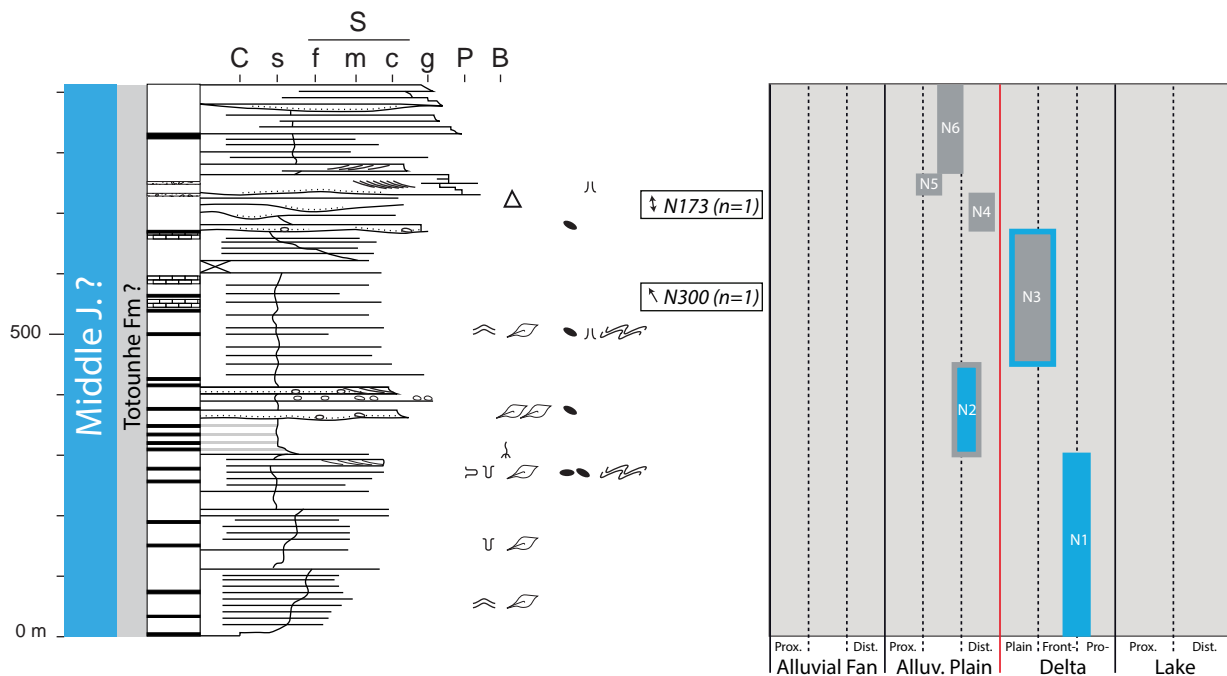


Figure 2.7.1: Log of the Mesozoic succession at the Nileke section in the Hexilagen Basin, on the southern flank of the North Tian Shan range, with paleocurrent directions, depositional environment interpretations (from alluvial plain to delta) and supposed climatic context. Scale 1/10 000.

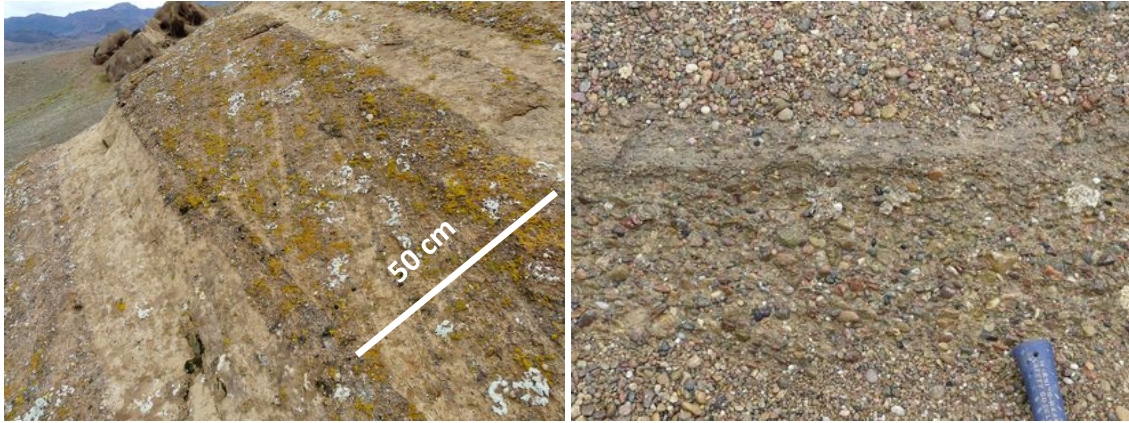


Figure 2.7.2 : Picture of the Cretaceous deposits in the Hexilagen Basin, close-by to the Nileke section. They show sandy channel deposits with megaripples (several decimetres) (a), and pebbly conglomerates (b).

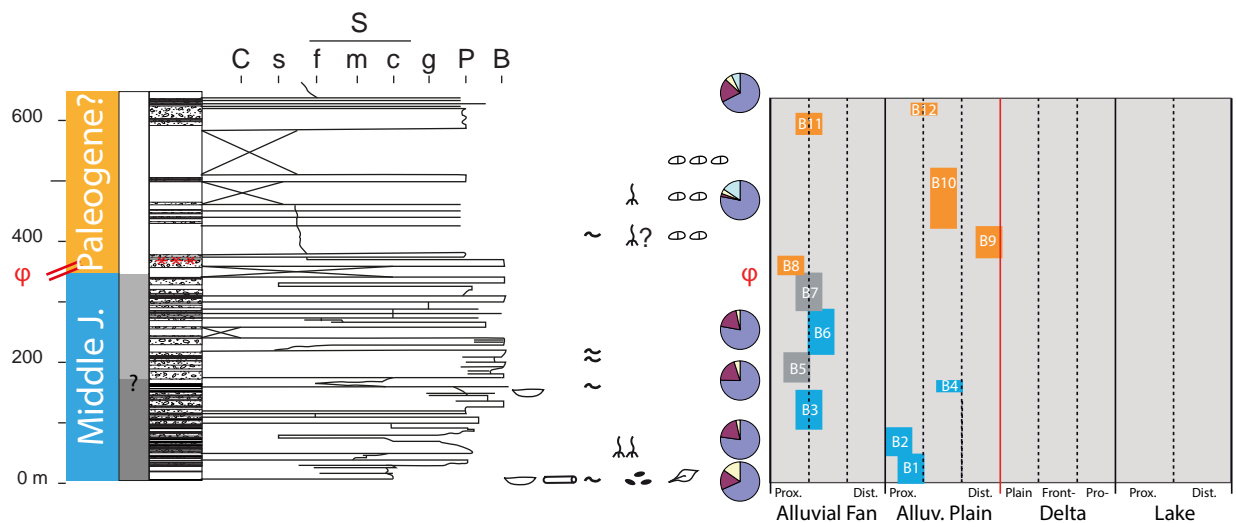


Figure 2.8 : Log of the Mesozoic succession at the Bayanbulak section, in the north of the Bayanbulak basin, with paleocurrent directions, depositional environment interpretations (from alluvial fan to alluvial plain). Scale 1/10 000.

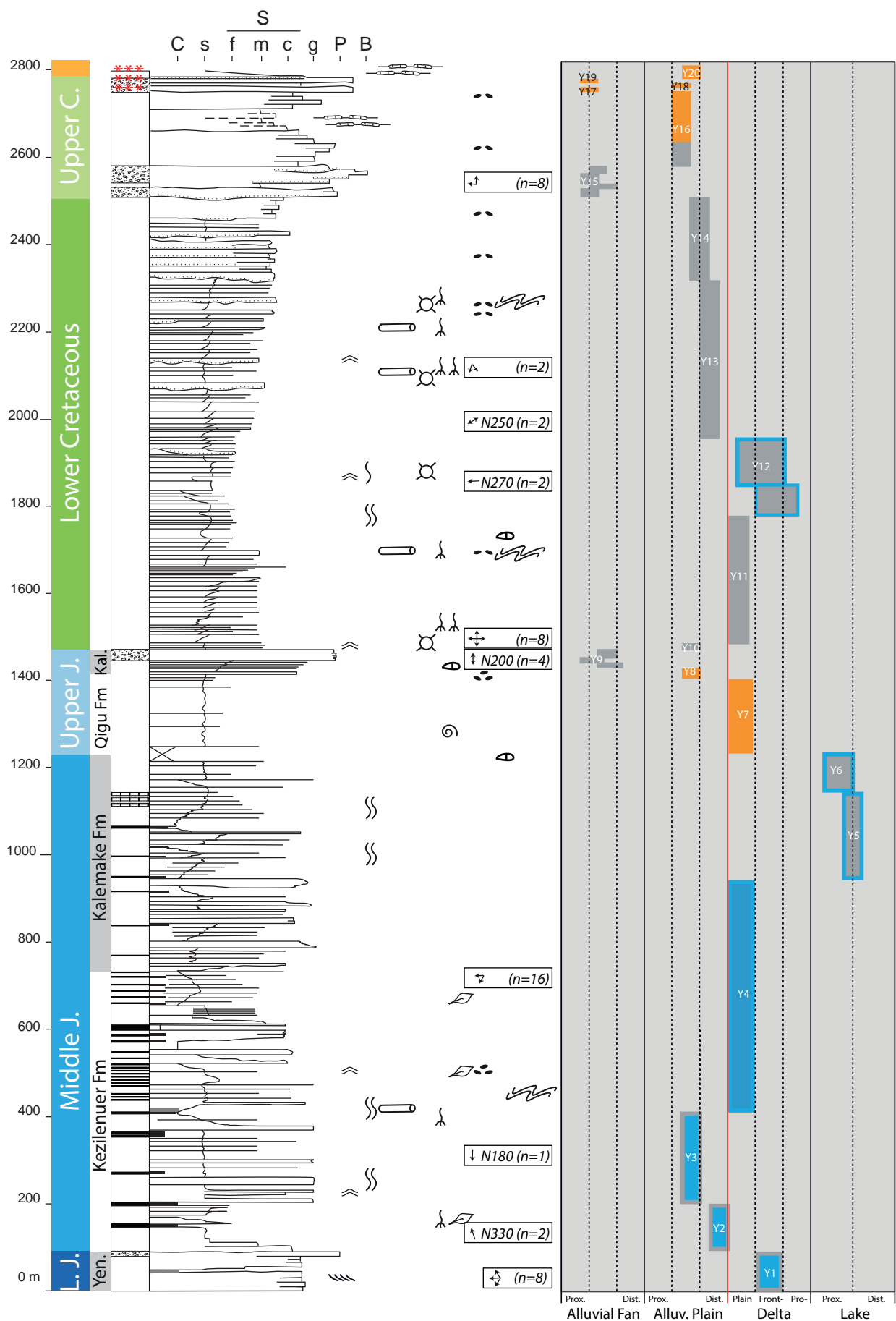


Figure 2.9: Log of the Mesozoic succession at the Yaha section, southern Tian Shan foothills, with paleocurrent directions, depositional environment interpretations (from alluvial fan to lake). Scale 1/10 000.

Fig. 2.10.1: Manas sectio



a. Unit M1 – Median alluvial plain (T3?); b. Units M4 et M5 – Proximal fan (J1); c. Unit M9 - Median alluvial plain (J1); d. Unit M10 - Prodelta and distal frontdelta (J1); e. Unit M11 - Median alluvial plain (J1); f. Unit M15 - Distal alluvial plain (J2).

Fig. 2.10.2: Manas sectio



a. Unit M16 - Deltaic plain (J2); b. Units M18, M19 & M20 - Pro- and frontdelta, deltaic plain (J2); c. Unit M21 & M22 – Distal and median alluvial plain (J2); d. Unit M22 – Distal-median alluvial plain (J2); e. Unit M24 - Prodelta (J3).

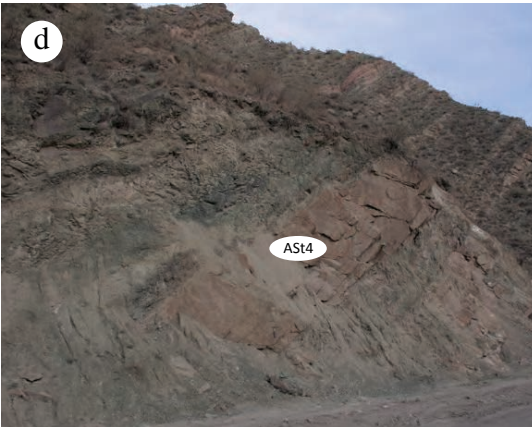
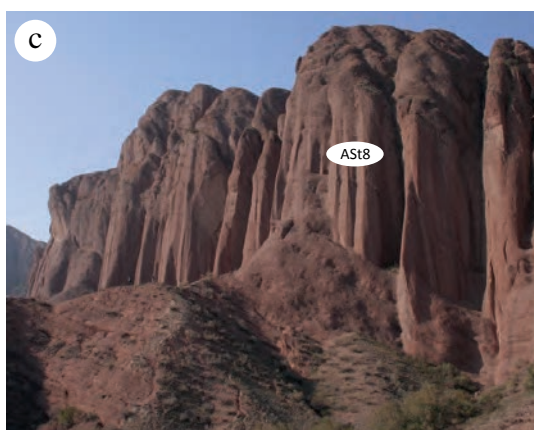
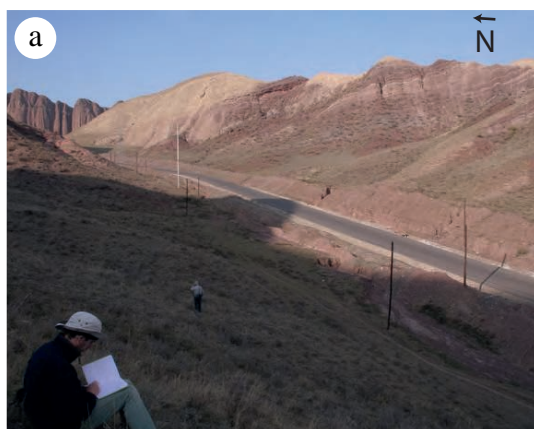


Fig. 2.10.3: Manas sectio



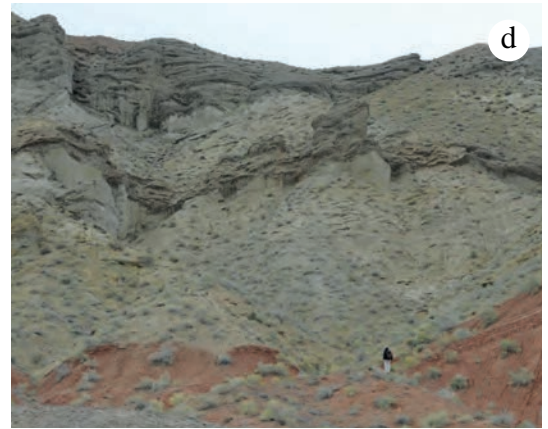
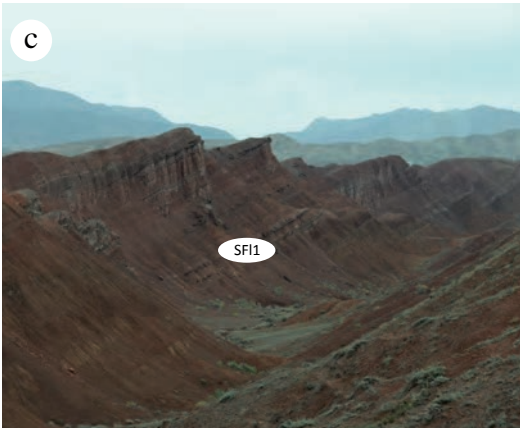
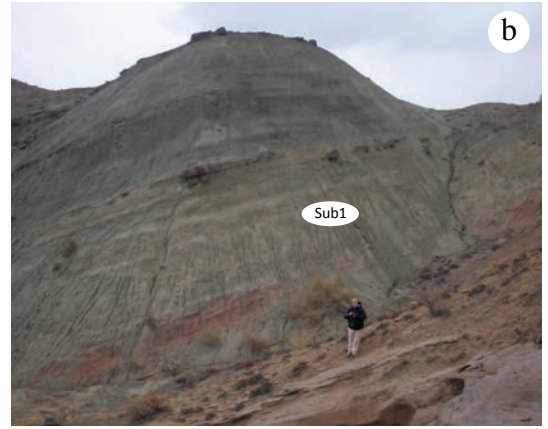
a. Unit M25 (and unit M29 in the background) - Distal alluvial plain (J3); b. Unit M26, M27, M28 & M29 - Distal alluvial plain to proximal alluvial fan (J3); c. Units M28 & M29 - Distal to proximal alluvial fan (J3); d. Unit M28 – Distal alluvial fan (J3); e. Units M30 & 31 - Lake (K1); f. Unit M32 – Median to proximal alluvial fan (K2).

Fig. 2.10.4: Manas sectio



a. & b. Unit M32 - Median to proximal alluvial fan (K2); c. Unit M32 (top) - Calcrete (Cret./Paleogene?); d. Unit M33 – Median alluvial plain (Paleogene).

Fig. 2.11: South Totoun section



a. Unit TS1 – Distal to median alluvial plain (J2); b. Unit TS2 - Delta (J2); c. Unit TS3 – Distal alluvial plain (J3); d. Unit TS6 - Median alluvial plain, with aeolian deposits (J3); e. Unit TS7 – Median alluvial fan (J3); f. Unit TS10 - Prodelta and proximal lake (K1).

Fig. 2.12: North Totoun section

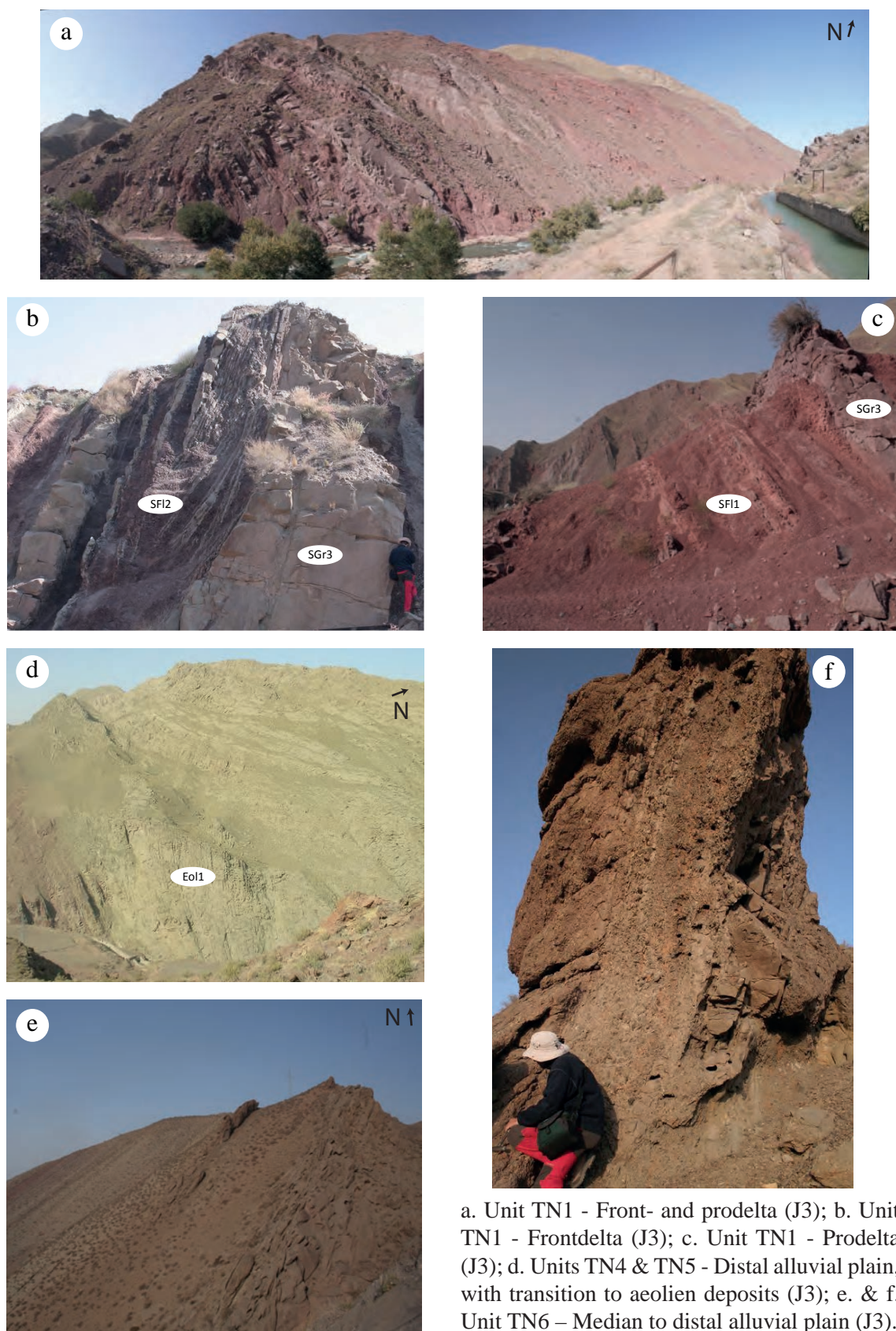
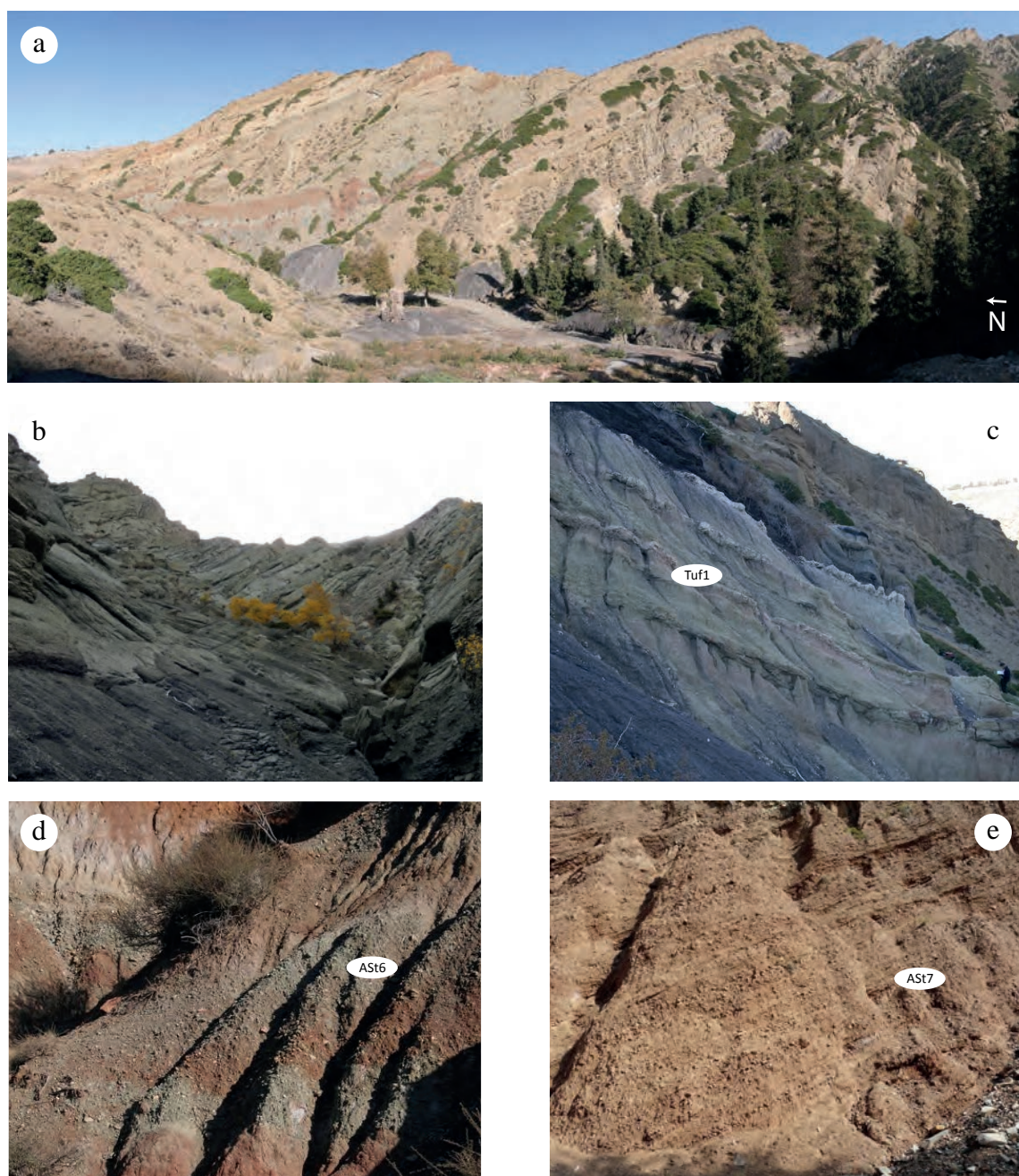
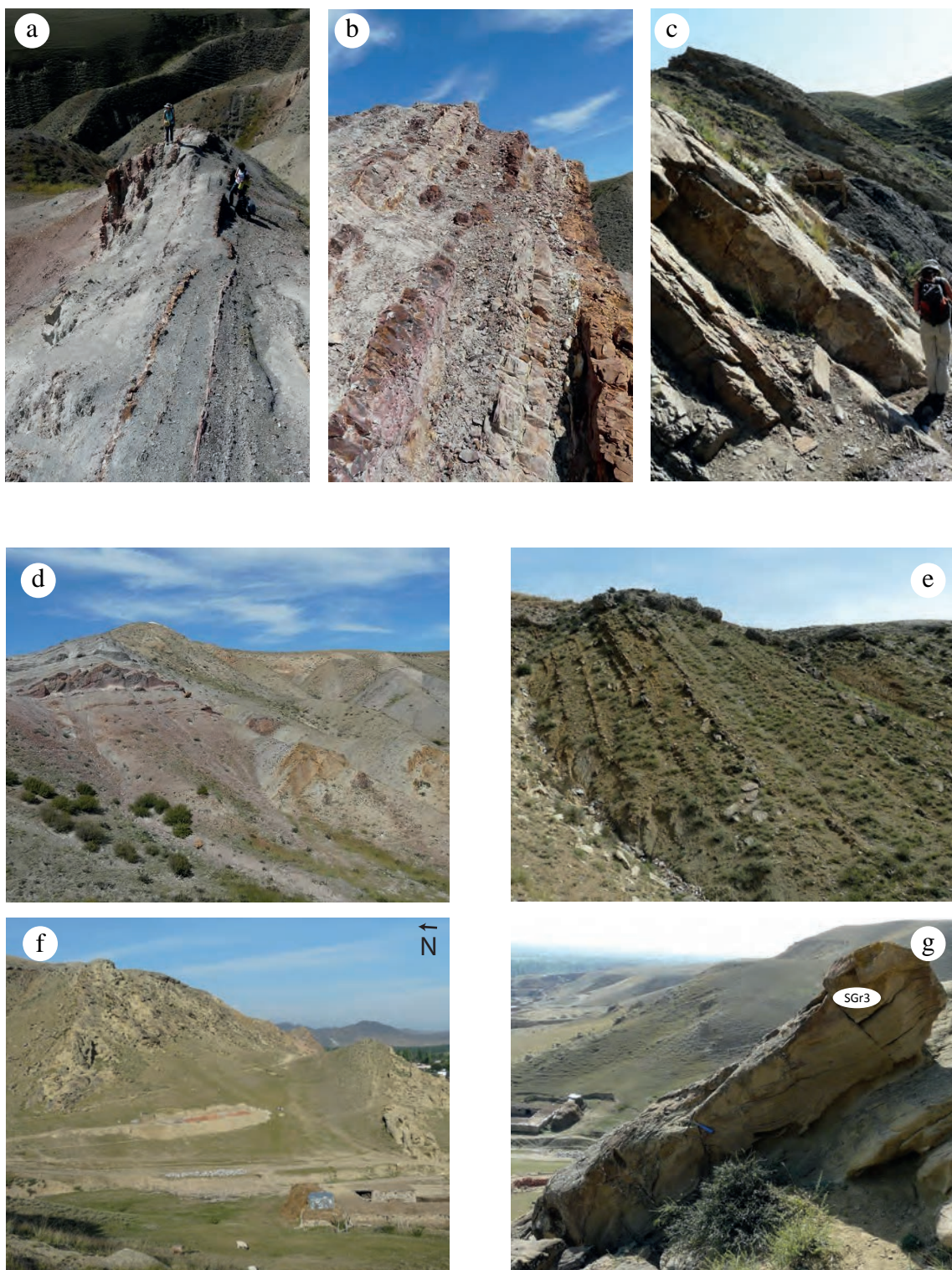


Fig. 2.13: Wusu section



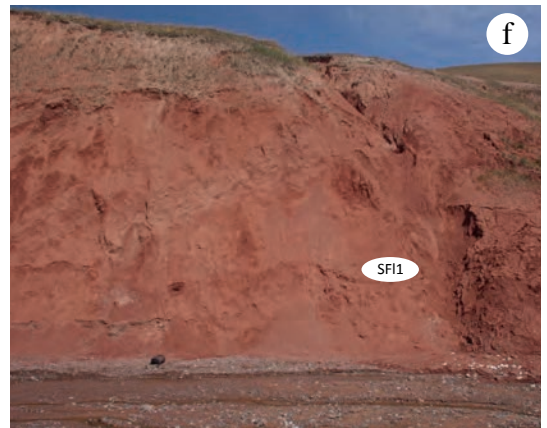
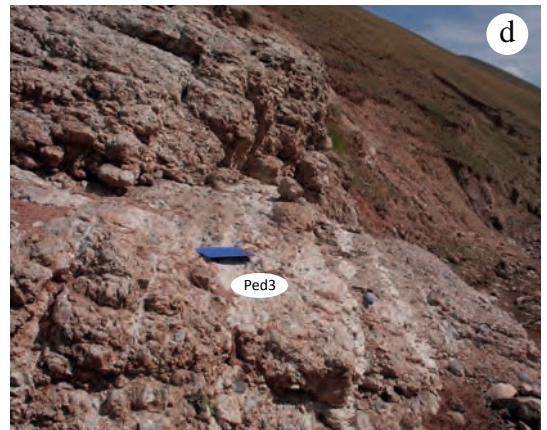
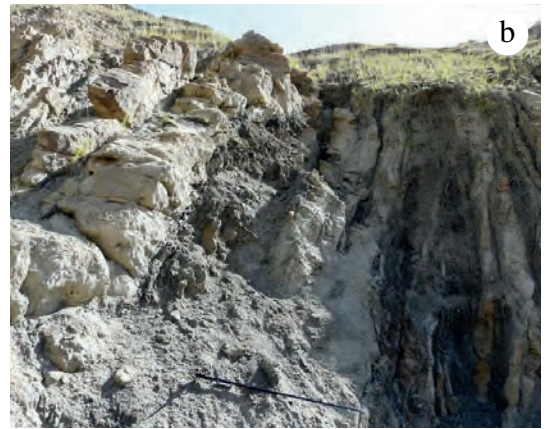
a. Overview of the lower part of the section (units W1-W12); b. Unit W1 – Proximal alluvial fan (J2); c. Unit W8 - Proximal alluvial plain (J2); d. Unit W10 - Distal alluvial fan (J2); e. Unit WE5 – Proximal-median alluvial fan (J3).

Fig. 2.14: Nileke sectio



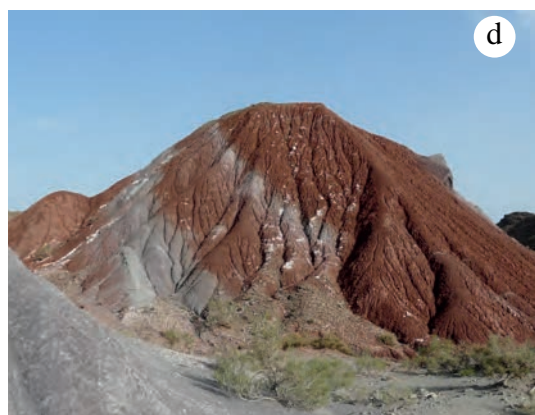
a. Unit N1 - Prodelta (J2); b. Unit N1 - Frontdelta (J2); c. Unit N1 - Frontdelta (J2); d. Unit N2 – Distal-median alluvial plain (J2); e. Unit N4 - Distal alluvial plain (J2); f. & g. Unit N5 - Median alluvial plain (J2).

Fig. 2.15: Bayanbulak



a. Unit B1 – Proximal alluvial plain (J2); b. Unit B4 - Median alluvial plain (J2); c. Unit B8 - Calcrete (Upper Cret./Tert.); d. Unit B8 - Calcrete (Upper Cret./Tert.); e. Unit B9 – Distal alluvial plain (Tert. ?); f. Unit B10 - Median alluvial plain (Tert. ?).

Fig. 2.16.1: Yaha section



a. Unit Y1 - Frontdelta (J1);
b. Unit Y2 - Distal alluvial plain (J2); c. Unit Y4 - Deltaic plain (J2); d. Unit Y7 - Deltaic plain (J2); e. Unit Y8-Y9 - Median alluvial plain - Fan (J3); f. Unit Y9 - Proximal alluvial fan; g. Unit Y11 - Distal alluvial plain (K1).

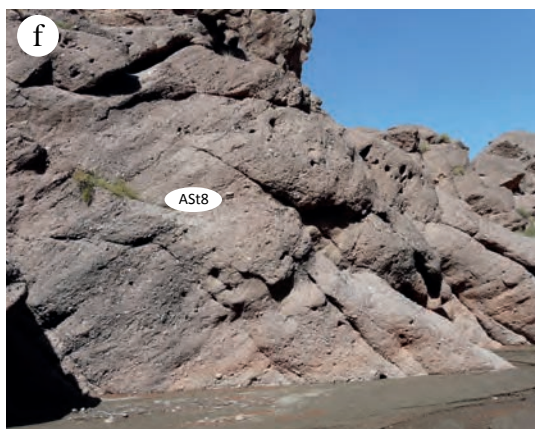


Fig. 2.16.2: Yaha sectio



a. Unit Y12 - Delta (K1); b. Unit Y13 – Distal alluvial plain (K1); c. Unit Y13 – Distal alluvial plain (K1); d. Unit Y14-Y15 – Distal-median alluvial plain - Fan (K1); e. Unit Y16 – Distal alluvial plain (K2); f. Unit Y17-20 - Calcrete and distal alluvial plain (Upper Cret./Tert. ?).

Figure 2.18

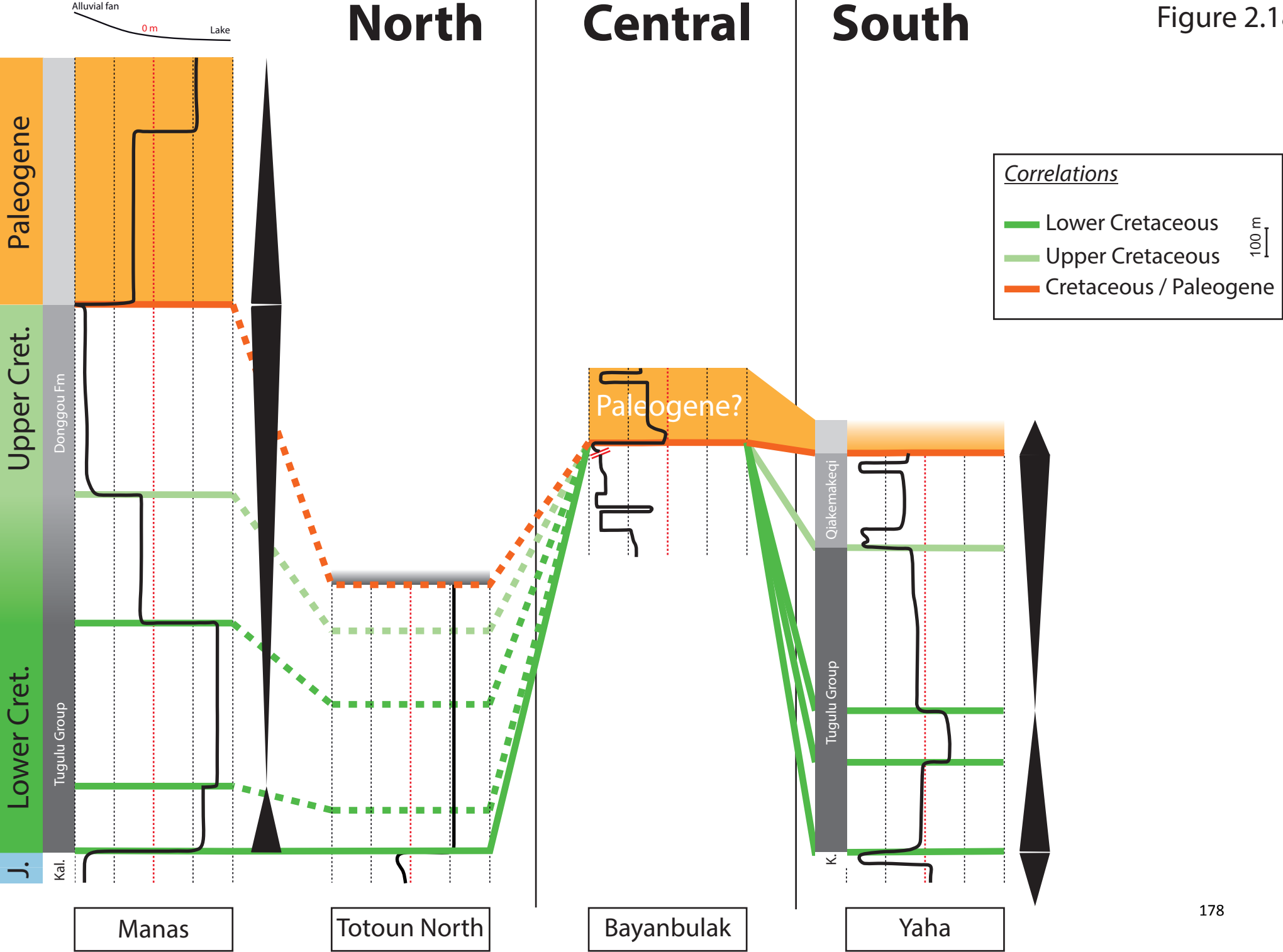
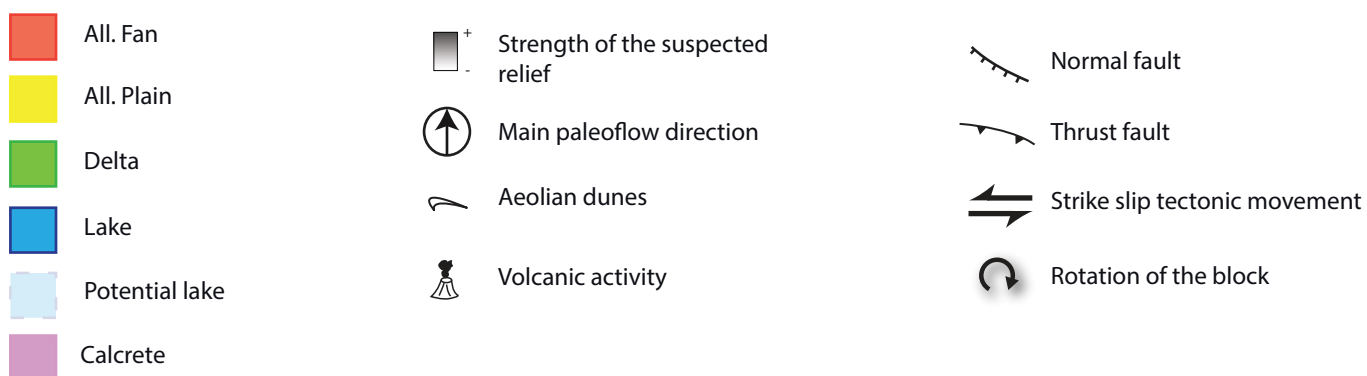
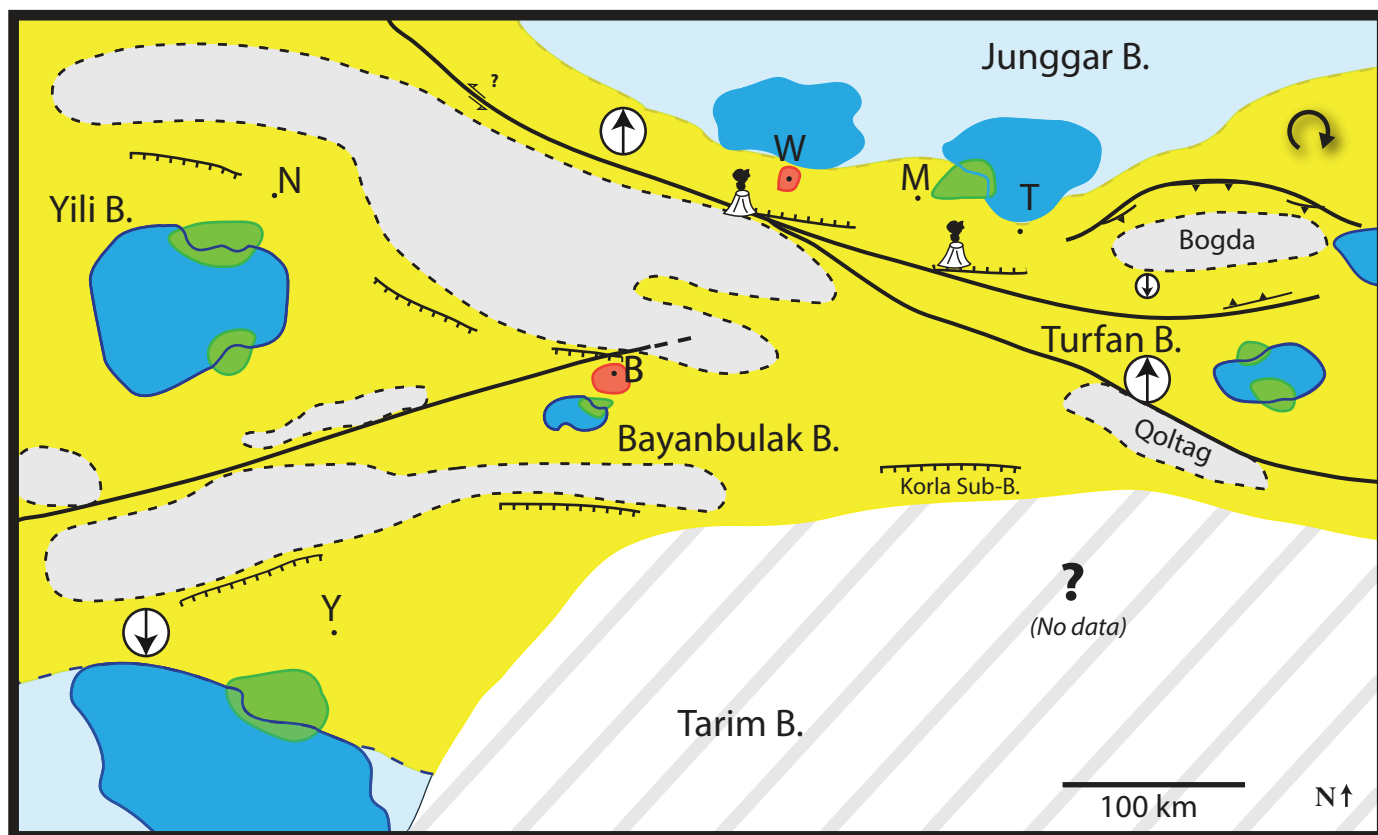
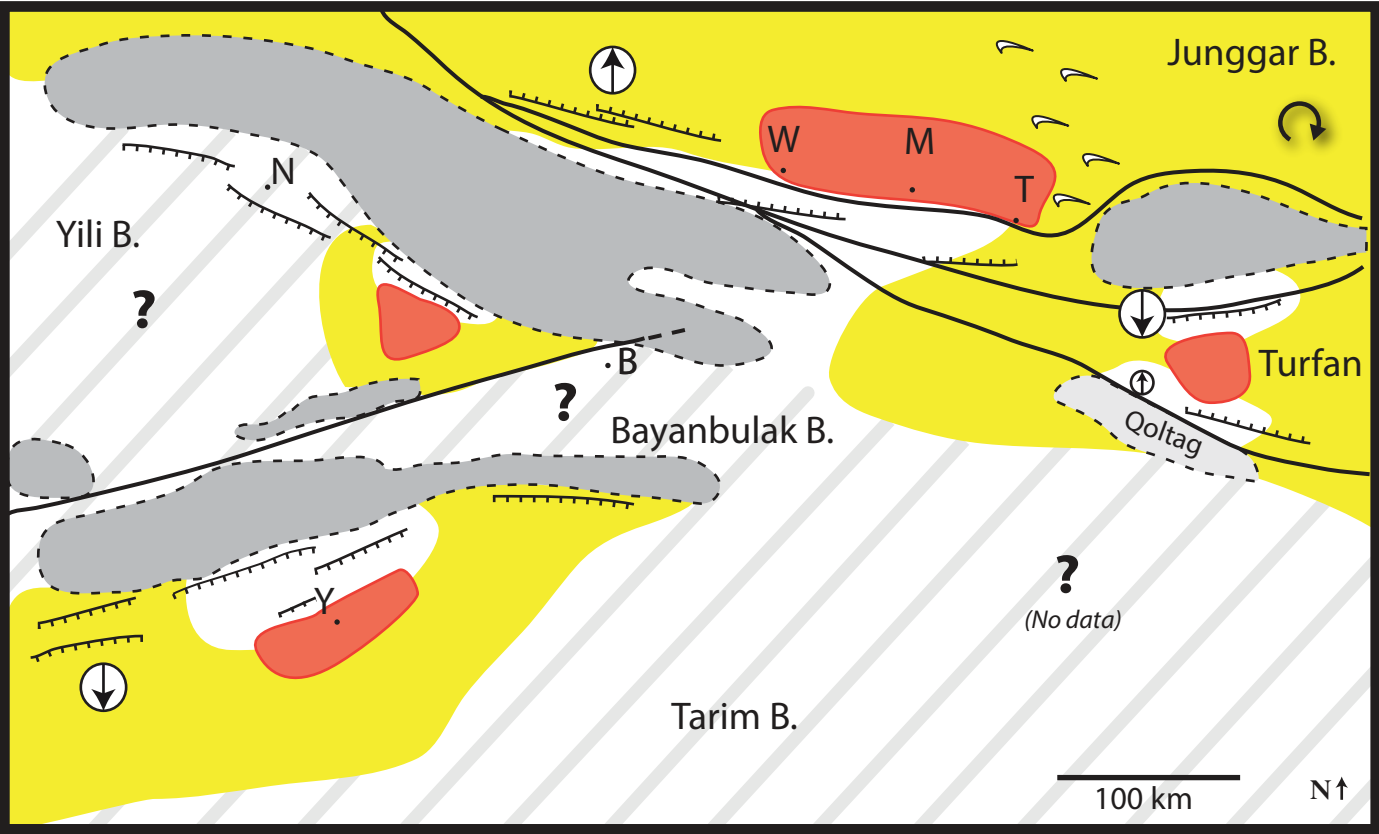


Figure 2.19: Paleoenvironmental reconstruction sketches of the Tian Shan area.

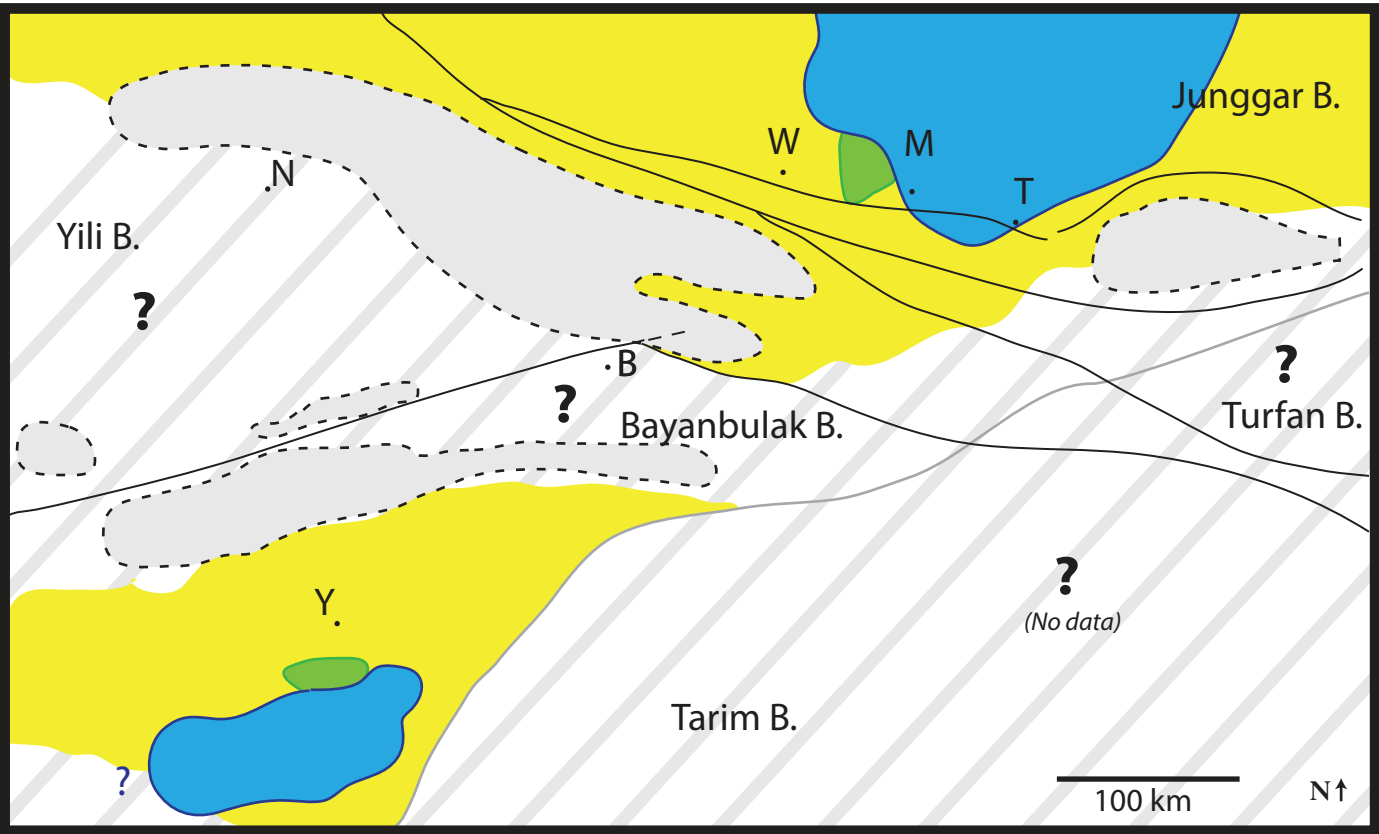
a] MFS? Early Qigu Fm (Middle Jurassic)



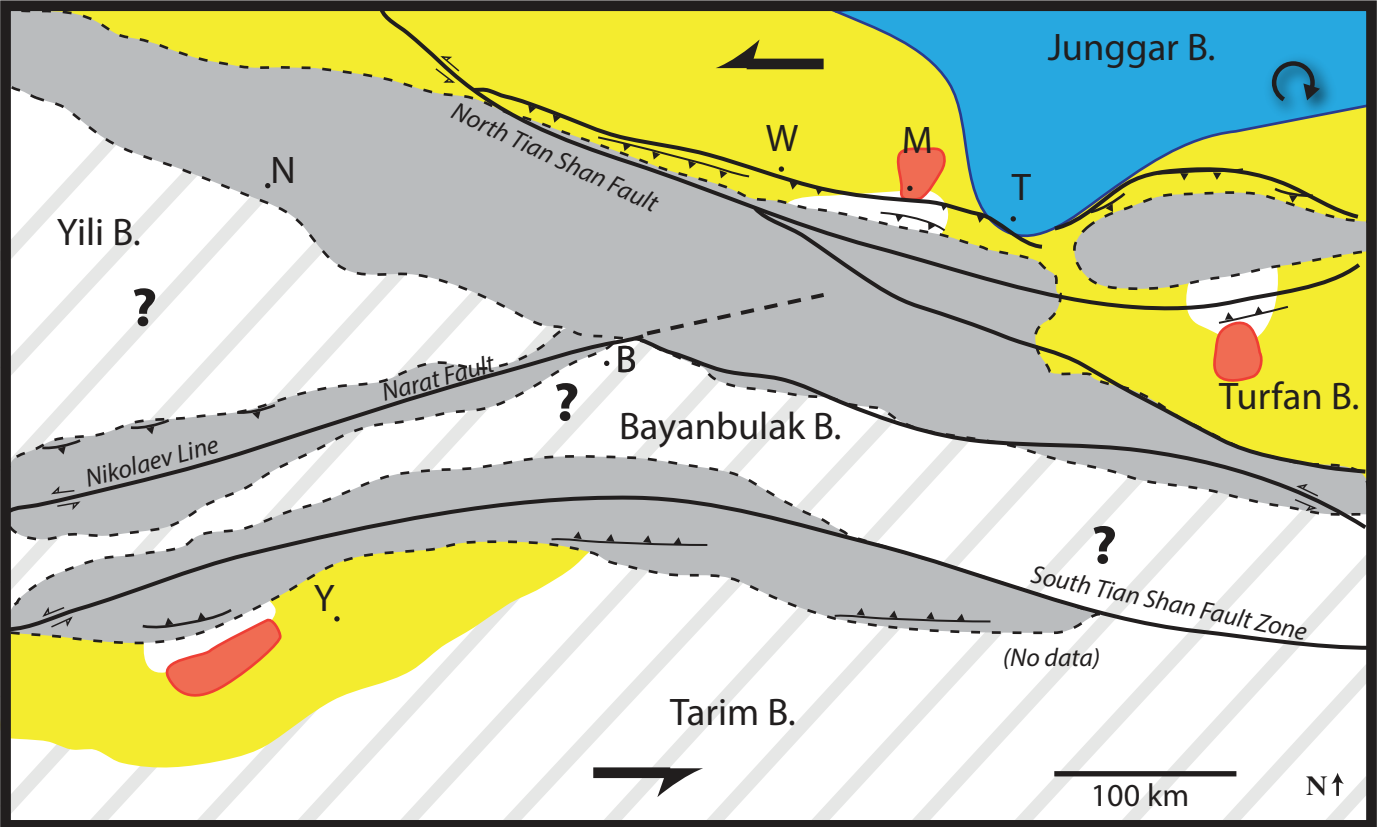
b) Kalazha Fm (Upper Jurassic)



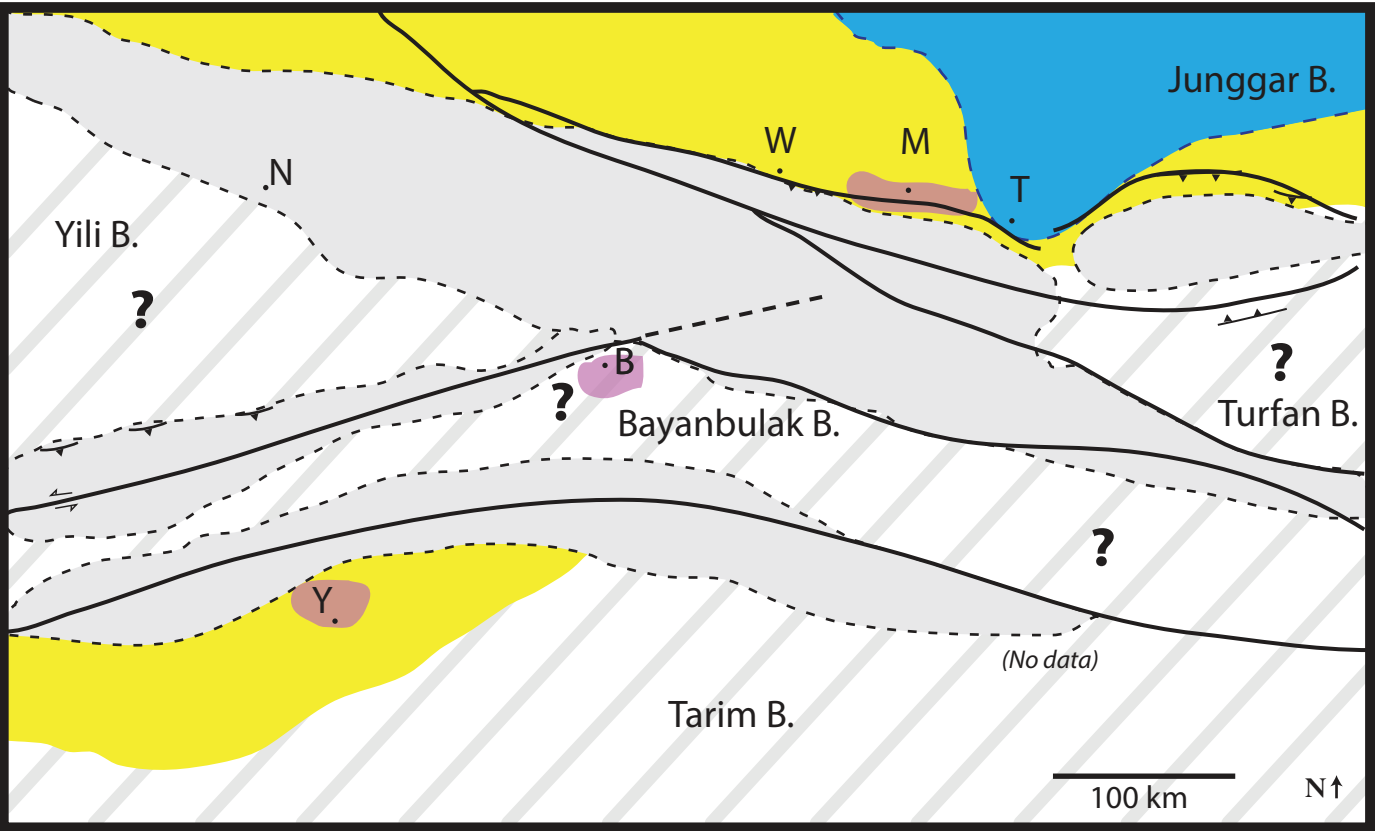
c) Lower Cretaceous



d] Upper Cretaceous



e] Early Paleogene



		Series	Formations					Series	Formations					Series	Formations										
		S-JUNGGAR										TURFAN							N-TARIM						
				Hendrix et al., 1992	Eberth et al., 2001			Ashraf et al., 2010			Hendrix et al., 1992			Hendrix et al., 1992	Wang et al., 2013										
Cretaceous		K2		Donggou (Aiolekehu)								Subashe			Qiakemakeqi										
								Kumutage																	
	Albian	K1	UNCONFORMITY		Lia.	?		UNCONFORMITY			?		UNCONFORMITY			Bashijiqlike									
	Aptian				She.																				
	Barremian				Hut.																				
	Hauterivian																								
	Valanginian		Tugulu Gr.	Lianmuqin	Qin.	?	Tugulu Gr.	Lianmuqin	Shanjinkou	Shusanhe	Baxigai	Baxigai													
	Berriasian			Hutubi				Sanshilidadun					Yageliemu	Yageliemu											
		Qingshuihe	Kal.	?																					
Jurassic	Tithonian	J3	Shishugou Gr.																						
	Kimmeridgian			Kalaza			Kal.																		
	Oxfordian			Qigu	Qig.																				
	Callovian	J2		Toutunhe	Tou.	?		Tou.																	
	Bathonian			Xishanyao	Xis.			Xis.																	
	Bajocian																								
	Aalenian																								
	Toarcian	J1			Sangonghe	San.		San.																	
	Pliensbachian				Badaowan	Bad.		Bad.																	
	Sinemurian																								
	Hettangian																								
	Triassic			T3	Xiaoquangou Gr.	Haojiagou			Haojia.																
		Huangshanjie				Huang.																			
		T2	Karamay																						
		T1	U. Cangfanggou Gr.																						

Heilbronn et al., in prep. - Table 1

Table 2 – Synthesis of the Mesozoic depositional environments in the Chinese Tian Shan area and supposed climate.

Syst.	Epoch	S- and NE-JUNGGAR	TURFAN-HAMI	N-TARIM	Climate
	Main references	Hendrix et al., 1992; Hendrix et al., 2000; Eberth et al., 2001 ; Vincent and Allen, 2001; Gao et al., 2003; Gao et al., 2004; Gao et al., 2009b; Hou et al., 2009; Bian et al., 2010 ; Hornung and Hinderer, 2011.	Hendrix et al., 1992; Shao et al., 1999; Greene et al., 2001; Shao et al., 2003.	Hendrix et al., 1992 ; Sobel, 1999; Jiang et al., 2008; Kuang et al., 2010; Wang et al., 2013.	Allen et al., 1991; Hendrix et al., 1992; Shao et al., 1999; Eberth et al., 2001 ; Thomas et al., 2001; Li et al., 2004 ; Tükten et al., 2004 cité dans Li 2004; Pfreundt et al., 2005; Sun and Wang, 2005; Jiang et al., 2008; Ashraf et al., 2010; Bian et al., 2010; Carroll et al., 2010; Sha et al., 2011; Pan et al., 2013.
Paleogene	Paleocene	Braid-delta and shallow lakes	Braided fluvial and alluvial plain to lacustrine, locally with salt-water-lakes		Arid-hot
			Unconformity (Hendrix et al., 1992; Shao et al., 1999)		
Cretaceous	Upp.	Alluvial, braided fluvial and braid-delta; caliche caps the unit	Braided fluvial to alluvial	Alluvial, with some caliche (Hendrix et al., 1992) OR braided delta, fandelta (Wang et al. 2013)	Arid
		Unconformity (Hendrix et al., 1992)	Unconformity (Hendrix et al., 1992)	Unconformity (Hendrix et al., 1992)	
	Low.	Fan-delta (basin margins), lacustrine (basin centre); fluvial lacustrine and eolian (NE basin-margin)	Braided fluvial – lacustrine delta (Shao et al., 1999) OR highly aggradational braided fluvial systems (Hendrix et al., 1992)	Lacustrine and delta (Wang et al., 2013); but no lacustrine environment in the Kuqa subbasin Hendrix et al. (1992)	Arid
		Local unconformity (Bian et al., 2010)	Unconformity (Shao et al., 1999)	Unconformity (Wang et al., 2013)	
Jurassic	Upp.	Alluvial plain to all. fan	Fluvial to alluvial	Deltaic plain or braided fluvial to alluvial (+debris flow)	Arid to semi-arid
	Mid.	Alluvial plain and lacustrine, delta, coal-deposits	Swamp, deltaic floodplain and meandering fluvial	Meandering fluvial, lacustrine and swampy facies	Warm and humid, seasonally dry
	Low.	Alluvial plain and shallow lacustrine	Fluvial delta and lacustrine	Meandering fluvial or front delta	Warm and humid ; from end of Lower Jurassic dry seasons
		Unconformity	Unconformity	Unconformity	
Triassic	Upp.	Retrogradation from all. fan to delta; first coal seams	Braided fluvial	Alluvial fan to lacustrine	Warm and humid
	Mid.	Alluvial fan and delta on the margins of the basin and large lakes in the center	Alluvial braided fluvial	Lacustrine (top)	From late Middle Trias: warm and humid
	Low.		Braided fluvial	T1, T2, T3 = alluvial, braided fluvial and lacustrine	Arid

TABLE 3 – Facies characteristics and interpretations in terms of depositional processes (see Photo Tables Fig. 2.2.1, 2.2.2, 2.2.3, 2.2.4 and 2.2.5).

Cal: calcareous production / Sub: subaqueous sediment fallout / Wav: wave-influenced deposition / SGr: subaqueous gravity processes / AGr: subaerial gravity processes / SFI: subaqueous flood inputs or sheet-flood / AFI: subaerial flood inputs or sheet-flood / ASr: subaerial stream-flow / Eol: Aeolian processes / OM: organic matter accumulation / Ped: pedogenic processes / Tuf: volcanic ash deposition .

In most cases, bioturbation stands for burrows.

NEW	Lithology	Bedding	Sedimentary structures	Miscellaneous	Processes	References
<i>Subaqueous deposits</i>						
Cal1	Cm to several dm-thick silty carbonate to calcareous siltstone or sandstone; light to dark grey or black	Homolithic		Occasional medium to strong bioturbation	Calcareous production mixed with a slight siliciclastic input	
Sub1	Several cm to several m-thick muddy to sandy siltstone; red, green, blue or gray	Homolithic	Massive to horizontal laminations	Occasional medium to strong bioturbation and gastropods	Subaqueous sediment fallout in a standing body of water	1, 2, 3, 4
Wav1	Cm to dm-thick fine to medium sandstone; red, green, blue or grey	Rare erosional basal boundaries	Symmetrical ripples; occasional silt-rich laminae	Occasional burrows on top of the bed	Fair-weather wave-influenced deposition	6, 7, 8, 9
SFI1	Cm to several m-thick heterolithics composed of (1) cm to several m-thick siltstone, locally clayey and (2) cm to several dm-thick fine sandstone; blue, grey, green, red or orange	Heterolithic	Massive or with horizontal laminations, local asymmetrical laminations	Local extra-size granule clasts; local light to strong bioturbation (on top of the laminae and beds) and rare gastropods	Deposition from subaqueous gravity flows alternating with sediment fallout ?	3, 5, 8, 10
SFI2	Several dm to several m-thick fine to coarse sandstone; white, yellow, orange or red	Heterolithic	Mega trough cross-bedding or horizontal planar bedding, sometimes underlined by coarse sand or granule deposits	Local extra-size granule clasts; local strong bioturbation within the cross-bedding	Deposition of a sandy bedload transported by subaerial to subaqueous unconfined flows (mouth bar)	5, 7, 11, 12, 9
SGr1	Cm to several dm-thick fine sandstone to granule conglomerate	Heterolithic; Occasional slight erosional basal boundaries	Sequences of normal grading, asymmetrical ripples to mega-ripples, and occasional symmetrical ripples		Gravity processes: from flooding induced processes to real turbiditic currents	e.g. 13, 14, 15, 4, 16 (real turbiditic currents)
SGr2	Dm to m-thick siltstone to coarse sandstone	Homolithic	Soft-sediment deformation	Mud rip-up clasts; locally bioturbation on top of the beds	Slump	
SGr3	Several dm to several m-thick fine to coarse sandstone; white, yellow, orange or red	Homolithic sandstone in continuous layers or locally in dm to m-wide lenses; occasional	Poorly to well sorted; flute casts on the base; local dewatering structures	Mud rip-up clasts; locally plant remains; weak to strong bioturbation (throughout the whole bed or just on	Gravity currents induced by flood events	14, 15, 17, 16

AGr1	Clast-supported going gradually into matrix-supported conglomerates, composed of subangular to subrounded pebbles; sandy matrix	erosional basal boundaries Meter to several m-thick beds; locally erosive basal boundary; poorly sorted; locally dm to m-wide sandy lenses	Upward increase of the matrix-content, locally graded; massive to faint planar bedding	top or base) Local wood remains	Subaerial to subaqueous debris flows	18, 15, 19, 20, 21, 22
<i>Subaerial deposits</i>						
Sub2	Several cm to several dm-thick siltstone; clayey to sandy; red, green, blue, gray or black; locally organic matter-rich or with several mm to cm-thick laminae	Homolithic	Massive to horizontal laminations	Occasional mudcracks; occasional root casts (facies Ped1); occasional plant remains	Subaqueous sediment fallout	3, 5
AFI1	Several cm to several dm-thick fine to coarse sandstone	Homolithic	Climbing ripples		Deposition of a sandy bedload transported by subaerial sheetflows (overbank)	23, 24
Eol1	Several dm to several m-thick fine to coarse sandstone		Well to moderately sorted; Sandflow and grainfall laminations, inverse climbing ripples in opposite directions down the forsets		Avalanches on dune leesides and grainfalls in zone of flow separation and wind ripples, subcritical climbing translational stratification	25, 26
Eol2	Fine to medium sandstone		Well to moderately sorted; Planar lamination or very-low angle cm-scale cross-laminations, inverse climbing ripples		Traction deposition by high wind velocity, aeolian sand sheet	25, 26, 27
Eol3	Fine to medium sandstone		Well sorted; Fine crude or irregular cm-horizontal lamination, adhesion plane beds, warts and ripples, inverse climbing ripples		Dry wind-blown sand to a wet or damp surface	25, 28, 29, 30
ASt1	(1) dm to several dm-thick homolithic with amalgamated cm to several cm-thick fine to coarse sandstones or (2) alternation of several cm to dm-thick sandstone with several cm to several dm-thick siltstone, and locally interbedded with coal		Asymmetrical ripples		Deposition of a sandy bedload transported by subaerial stream- or sheet-flows	2
ASt2	Several cm to several dm-thick heterolithic of (1) siltstone to fine sandstone and (2)	Locally m to several m-thick thinning-out or	Moderately sorted, occasional normal grading; planar	Occasional mud rip-up clasts; plant remains; occasional	Multiple stacked deposition of sandy bedloads transported	2

	fine to coarse sandstone; green, grey, black, rust-coloured, orange, red, purple, brown; locally organic matter-rich layers, coaly sandstones or coal chips	thickening-up heterolithic sequences; locally m to several m-wide lenticular beds	laminations or asymmetrical ripples	root casts (facies Ped1)	by subaerial stream- and sheet-flows	
ASt3	Several dm to several m-thick fine to coarse sandstone and small pebbles; beige, green-grey or red	Sandstone in continuous layers or locally in m to several m-wide lenticular beds	Poorly sorted sandstone; trough cross-bedding to mega-ripples; amalgamated events; sometimes erosional basal boundaries; foresets can also be underlined by granule to pebble lags; mud rip-up clasts	Plant remains; rare root casts (facies Ped1); locally with several dm-thick pebble lags or with floating pebbles (with one-gravel-thick lag deposits)	Deposition of a sandy bedload transported by subaerial stream-flows	31, 1, 2, 32, 23
ASt4	Several dm to several m-thick, mostly fine to medium sandstone, and locally siltstone or coarse sandstone with some granules	Homolithic; Continuous layers or locally dm to several m-wide lenticular beds	Poorly sorted; succession of stacked planar and trough cross-bedding, sometimes overlapping; erosional basal boundary; parting lineation; rare to abundant mud rip-up clasts; foresets can also be underlined by granule to pebble lags; can also be massive		Deposition of a sandy bedload transported by subaerial stream-flows	31, 1, 2, 32, 23
ASt5	Several dm to several m-thick fine to coarse sandstone	Homolithic	Planar laminations		Deposition of a sandy bedload transported by subaerial stream-flows	31, 1, 2, 32, 23
ASt6	Clast-supported conglomerates composed of subangular to subrounded pebbles; locally more matrix-rich (sandy); locally coal or silt- to sandstone lenses	Several dm to m-thick continuous layers or several m-wide lenticular beds; locally dm to several dm-thick sandy interbeds; locally erosional basal boundaries	Poorly sorted; massive to mega-ripples; locally winnowed surfaces	Locally associated to flooding facies (OM, Sub2, ASt1)	Deposition of a sandy to gravely bedload transported by subaerial stream-flows	31, 1, 2, 32, 33, 23
ASt7	Clast-supported conglomerates composed of subangular to subrounded pebbles to cobbles; locally matrix-rich (silt to coarse sand) or with lenticular beds (silt- to coarse sandstone); locally reworked coal-chips	Meter to tens of m-thick beds, locally dm to m-wide sandy lenses; occasional erosional basal boundaries	Poorly sorted; locally graded; faint planar bedding or faint low-angle cross-bedding; pebble imbrications; locally erosional basal boundary	Locally reworked coal fragments [X mm]; locally barely burned plants fragments	Deposition of a gravelly bedload transported by subaerial stream- or sheet-flows	31, 1, 2, 32, 33, 23

ASt8	Clast-supported conglomerates composed of angular to subrounded pebbles to cobbles; locally matrix-rich (silt to coarse sand) or with lenticular beds (silt- to coarse sandstone)	Meter to tens of m-thick beds, locally dm to m-wide sandy lenticular beds	Poorly sorted; locally graded; faint planar bedding or faint low-angle cross-bedding; pebble imbrications; locally winnowed surfaces	Locally root casts (facies Ped1)	Deposition of a gravelly bedload transported by subaerial temporary stream- or sheet-flows (until flash flood ?)	31, 1, 2, 32, 33, 23
AGr2	Matrix-supported conglomerates composed of subangular to subrounded granules to cobbles; sandy matrix, sometimes with small gravels	Several dm to several m-thick beds; poorly sorted	Massive		Subaerial debris flow	34, 15, 35, 20, 21
OM	Organic matter-rich mud- and siltstone or sandstone, locally until coal deposits	Cm to dm-thick continuous layers or locally dm to m-wide lenticular beds	Plant remains and root casts		Organic-matter accumulation	

Pedogenetic features

Ped1	Cm to several dm-long root casts		Perpendicular to the bedding	Locally with an oxydation-halo	Short-term soil-formation	36
Ped2	Loose calcareous nodules ("Caliches") to coalescent beds	From poorly to strongly coalescent (cm - X dm)			Semi-arid pedogenic processes	36, 37
Ped3	Massive, several m-thick pebble beds, with coarse sandstone-matrix and pervasive calcareous impregnation/cementation ?	From nodular and powdery to highly indurated and massive (dm - X m)	Local continuous calcareous-coating on pebbles		Long lasting wet-dry pedogenic processes	36, 37

Others

Tuf1	Fine white sandstone with small granules	Massive; several cm to several dm-thick laterally continuous layers	Locally with floatting granules	Locally plants remains	Subaerien volcanic ash deposition	
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1) Miall, 1977; 2) Miall, 1978; 3) Hyne et al., 1979; 4) Horton and Schmitt, 1996; 5) Tye and Coleman, 1989; 6) Chan and Dott, 1986; 7) McCormick and Grotzinger, 1993; 8) Plint and Brown, 1994; 9) Buatois et al., 2012; 10) Helland-Hansen, 2010; 11) Marshall, 2000; 12) Plink-Björklund and Steel, 2005; 13) Bouma, 1962; 14) Middleton and Hampton, 1976; 15) Lowe, 1982; 16) Mulder and Alexander, 2001; 17) Kneller and Branney, 1995; 18) Walker, 1975; 19) Nemec and Steel, 1984; 20) Smith, 1986; 21) Sohn et al., 1999; 22) Sohn, 2002; 23) Mack and Leeder, 1999; 24) Hampton and Horton, 2007; 25) Hunter, 1977; 26) Clemmensen & Abrahamsen, 1983; 27) Brookfield & Silvestro, 2010; 28) Kocurek & Fielder, 1982; 29) Kocurek & Nielson, 1986; 30) Brookfield, 1992; 31) Simons et al., 1965; 32) Southard, 1991; 33) Nemec and Postma, 1993; 34) Major, 1997; 35) Nemec and Steel, 1984; 36) Retallack, 1988; 37) Alonso-Zarza, 2003.

TABLE 4 – The main depositional environments of the Mesozoic series described in the Tian Shan area.

<i>Depositional environment</i>	<i>Facies</i>	<i>Sedimentary architecture</i>
Lake Distal lake (L1)	Sub1, SGr1, Cal1, OM	Thick (dm to several m) siltstone deposits, dominated by Sub1, locally OM-rich, very few turbiditic events (cm to several dm-thick) and few silty carbonate beds (cm to several dm-thick)
Proximal lake (L2)	Sub1, SGr1, Cal1, Wav1, OM	Thick (dm to several m) siltstone deposits (Sub1, dominant) and sandstone beds (cm to several dm-thick) with erosive basal boundaries and locally with symmetric ripples (Wav1) and turbiditic events (dm-thick) (SGr1)
Delta Prodelta (D1)	Sub1, SGr1, Cal1, Wav1, SGr3, SGr2, SFI1, OM	Siltstone deposits (dm to several m, dominant) and heterolithic sandstone beds with erosive basal boundaries (dm to several dm-thick), locally with symmetric or asymmetric ripples; local dm to m-wide homolithic sandstone lenses (SGr3)
Frontdelta (D2)	SFI2, SGr2, Sub1, SGr1, Wav1, Ast1, Ast2, Ast3, OM	Thick monolithic sandstone deposits (several dm to several m) with ripples to megaripples; no channel; rare erosional basal boundaries; rare siltstone deposits (cm to dm)
Deltaic plain (D3)	ASt3, ASt4, ASt5, ASt6, AFI1, ASt1, Sub2, Sub1, SGr1, Wav1, Ped2, Ped1, OM	Thick (dm to several m) mostly through cross-bedded sandstones as channel-filling; locally amalgamated homolithic sandstones (dm to several m-wide) or heterolithic (several dm to several m) silt- and sandstones; locally clast-supported conglomerates (ASt6); locally symmetrical ripples, bioturbation or pedogenic features (Ped 1 and 2)
Alluvial Plain Distal (AP1)	ASt3, ASt4, ASt5, Ast1, Ast2, AFI1, Sub2, Ped2, Ped 1, OM Eol1, Eol2, Eol3	Heterolithic siltstone and fine sandstone deposits (cm to m-thick) (dominant) with sand-filled channels (dm to m-thick), with asymmetric ripples and megaripples. Siltstones are locally associated with coal or pedogenic features. Locally aeolian deposits
Median (AP2)	ASt3, ASt4, ASt5, ASt6, AGr2, Ast1, Ast2, AFI1, Sub2, Ped2, Ped1, OM	Heterolithic siltstone and sandstone deposits (cm to several dm-thick) with sand-filled channels (dm to several dm-thick) sometimes gravelly, containing asymmetric megaripples or horizontal planar bedding. Locally associated with coal or pedogenic features
Proximal (AP3)	ASt4, ASt5, ASt6, ASt3, AGr2, Ast1, Ast2, Sub2, Ped2, Ped1, OM	Heterolithic siltstone and sandstone deposits (dm to several dm-thick) containing asymmetric megaripples or horizontal planar bedding, with gravelly channels (dm to several dm-thick). Locally associated with coal or pedogenic features
Alluvial Fan Distal (AF1)	ASt7, ASt8, ASt4, ASt5, ASt6, AGr1	Clast-supported conglomerates (m to tens of m-thick) (dominant) and sandstones (dm to m-thick) containing asymmetric megaripples or horizontal planar bedding.
Median (AF2)	ASt7, ASt8, AGr1, AGr2	Clast-supported conglomerates with faint horizontal planar bedding or faint low-angle cross-bedding, locally with erosional basal boundary (m to tens of m-thick) (dominant), associated to matrix-supported conglomerates and sandstones (dm to m-thick)
Proximal (AF3)	AGr1, AGr2, ASt7, ASt8	Matrix-supported conglomerates either massive or with faint horizontal planar bedding (several dm to several m-thick) (dominant) and clast-supported conglomerates (m to tens of m-thick)

TABLE 5 – Description of the sections described in the Tian Shan area (see Fig. 2.1 for location, as well as Fig. 2.3, 2.4, 2.5, 2.6, 2.7, 2.8 and 2.9 for the logs).

The stratigraphic ages correspond to these from the Geological map (REF). – Miscellaneous: when not specified, bioturbations correspond to undetermined burrow features. – Facies listed here are not exhaustive, just the dominant ones (see Table 3). – Granulometry mentions the most dominant one first. – General depositional environment (see Table 4).

T3 = Upper Trias; J1 = Lower Jurassic; J2 = Middle Jurassic; J3 = Upper Jurassic; K1 = Lower Cretaceous; K2 = Upper Cretaceous; Cen. = Cenozoic.

Section	Unit	Stratigraphic age	Thickness (m)	General color / Photos	Miscellaneous	Facies	Granulometry/ Lithology	General Environment
Manas	M1	T3?	ca. 140	Grey, brown ; Fig. 2.10.1	Sharp and erosive M1-M2 boundary	ASt4, Sub2, AGr2	Sand, silt, pebbles	Proximal part of AP2
	M2	T3?	39	Grey, brown		ASt7, ASt4	Pebbles, coarse sand	AF2-AF1
	M3	T3/J1 ?	ca. 120	Grey, brown	Outcrop gap	Sub2	Silt	AP
	M4	J1	ca. 35	Grey, brown ; Fig. 2.10.1		ASt7	Pebbles and cobbles	AF3
	M5	J1	ca. 65	Grey, brown, black ; Fig. 2.10.1		ASt7, ASt3 Sub2 + OM	Sand	AF2-AF1, locally AP
	M6	J1	ca. 120?	Grey, brown	Poorly exposed ; M6-M7=erosive	ASt3, Sub2, ASt1, ASt6	Fine sand	AP1
	M7	J1	ca. 40	Grey, brown		AGr2, ASt7 + OM	Coarse sand with pebbles	AF, going until the most distal part of the fan
	M8	J1	190	Grey, black	Poorly exposed	ASt7, ASt4, ASt3 + OM	Pebbles	AF2, locally AP
	M9	J1		Grey, brown, black ; Fig. 2.10.1		ASt3, AGr2	Sand, pebbles	AP2
	M10	J1	ca. 160	Beige ; Fig. 2.10.1	Poorly exposed	Sub1, SFl1 + OM	Silt, sand	D1 (from L2 to distal D2)
	M11	J1	ca. 80	Beige ;		ASt4, ASt6	Sand, pebbles	AP2

			Fig. 2.10.1				
M12	J1	ca. 250	Beige		ASt4	Sand	AP2-AP1
M13	J1/J2 ?	763	Beige, black	Outcrop gap contains J1-J2 boundary ; M13-M14 = slightly erosional boundary	ASt2, ASt1, Sub2, Wav1, SGr2 + OM	Very fine to coarse sand, silt, pebbles	D3
M14	J2		Beige, black	M14-M15 = slightly erosional boundary	ASt6, ASt4 + OM	Pebbles, sand	AP3-AP2
M15	J2	66	Beige ; Fig. 2.10.1		ASt4, Sub2 + OM	Sand, silt	AP1
M16	J2	ca. 480	Grey, beige, black ; Fig. 2.10.2	Not continuously exposed	Sub2, ASt4 + OM	Silt, sand	D3
M17	J2	40	Grey, beige	Non-erosive boundary	Sub2, SGr3, SGr1 + OM	Silt, sand, pebbles	D2
M18	J2	150	Grey, black ; Fig. 2.10.2	M18-M19 = sharp and erosive boundary	Sub2, SGr3, Wav1	Sand, silt	D1
M19	J2	174	Beige, black, grey; Fig. 2.10.2		SGr3, SFl2, Sub2 + OM	Sand, silt	D2 (with some trends to D1)
M20	J2	ca. 50 (expected)	Grey, black, beige; Fig. 2.10.2		Sub2, SGr3/ASt4 + OM	Silt, sand	D3
M21	J2	ca. 440 (expected)	Beige, grey, black; Fig. 2.10.2	Tree stump	Sub2, ASt4 + OM	Sand, silt	AP1
M22	J2	ca. 575 (expected)	Red and purple, sometimes green or blue; Fig. 2.10.2		ASt4, Sub2	Sand, silt	AP1-AP2
M23	J2	164	Green, sometimes pink		SGr3, Wav1	Silt, sand	D2
M24	J3?	167	Grey, purple,		SFl1	Silt, sand	D1

				pink; Fig. 2.10.2				
	M25	J3	170	Red; Fig. 2.10.3		ASt4	Sand	AP1
	M26	J3	ca. 335	Red; Fig. 2.10.3	Sharp and erosive boundary (M26-M27)	SFl1, Sub2	Clayey silt, sand	Very distal AP (edge to D3)
	M27	J3	ca. 60	Red, purple ; Fig. 2.10.3		ASt2	Silt, sand	AP1
	M28	J3	61	Orange, beige ; Fig. 2.10.3		ASt2, ASt6 + Ped1	Silt, sand (clayey at the base and shift to poorly sorted pebble-conglomerates).	AF1 (most distal part)
	M29	J3	ca. 300	Red, brown ; Fig. 2.10.2, 2.10.3	M29-M30 = sharp and non-erosive	ASt8	Pebbles to cobbles, sand	AF3
	M30	K1	ca. 220	Grey, beige ; Fig. 2.10.3		Sub1, SGr1	Silt, sand, pebbles	L2
	M31	K1	ca. 550 + 430 m of gap	Dark grey ; Fig. 2.10.3	Outcrop gap (430 m thick)	Sub1, Wav1	Clay- and silt	L1
	M32	K2	650	Red, grey; Fig. 2.10.3, 2.10.4	M39-M33 = sharp and non-erosive	AGr2, ASt6 + Ped3 (last eight metres)	Conglomerate (pebbles to cobbles), sand	Most proximal part of a AF2(local ly proximal)
	M33	Ceno z.	590	Red, blue ; Fig. 2.10.4	Poorly exposed	Sub2 + Ped2	Clay- or silt	AP2
	M34	Ceno z.	> 240	Orange, beige		Wav1, Sub 1	Fine sand- and silts, locally clayey or medium sand to pebbles	L2
South Totoun	TS1	J2	431	Yellow, beige ; Fig. 2.11	TS1/TS2 = sharp and non-erosive	ASt4, ASt3 + OM	Silt, medium sand, few small pebbles	AP1-AP2
	TS2	J2/J3	514	Red, purple, grey, blue; Fig. 2.11	Commonly bioturbated; TS2/TS3 = sharp and non-erosive	Sub1, SGr3, SGr2, AFl1	Silty claystone, silt, sand	D (D2 to D1, and D3)
	TS3	J3	234	Red,	Commonly	SFl1	Silt, sand	AP1

				orange, purple; Fig. 2.11	bioturbated; TS3/TS4 = not exposed	+ Ped2, Ped1		
	TS4	J3	95	Orange, red	TS4/TS5 = sharp and non-erosive	SFI1, SFI2	Silt, sand	D3
	TS5	J3	ca. 190	Orange	TS5/TS6 = sharp	ASt4, ASt3 Eol2 + Ped1	Sand	AP1-AP2
	TS6	J3	ca. 50	Green-beige, orange; Fig. 2.11		Eol1, Eol2, Eol3	Sand, silt	AP2-AP1
	TS7	J3	28	Grey, orange-brown, yellow; Fig. 2.11	Sharp, basal erosive boundary	AGr2, ASt3, ASt6	Pebbles to cobbles, sand	AF2
	TS8	J3	53	Grey, yellow	Locally winnowed sand	ASt3	Sand, pebbles	AP2-AP1
	TS9	J3	32	Grey, yellow	Sharp, basal erosive boundary	AGr2, ASt3, ASt6, ASt4	Pebbles to cobbles, sand	AF2, with a tendency to AF3
	TS10	K1	338	Grey, yellow-orange-purple-blue; Fig. 2.11	Locally calcareous silts, locally bioturbated	SGr1, SFI2, SGr3, AFI1, ASt1, Wav1	Silt, sand, rare conglomerates at base (granules)	D1 to L2
North Totoun	TN1	J3	657	Red, beige; Fig. 2.12	Locally strongly bioturbated (horizontal and vertical burrows); TN1/TN2 = sharp and probably erosive	SFI1, SGr3, SFI2, SGr2 + Tuf1?	Sand, silt, few pebbles	D1 to D2
	TN2	J3	63	Red	Horizontal and vertical burrows; TN2/TN3 = sharp and non-erosive	ASt3, ASt4, ASt5, Sub2, Wav1 (rare)	Sand, silt	AP2 (locally distal)
	TN3	J3	21	Grey, red	Strongly bioturbated	SGr3, Sub2, Wav1	Sand, silt	D3
	TN4	J3	19	Red, grey; Fig. 2.12	TN4/TN5 = erosive unconformity	Eol2, Eol1	Sand	AP2-AP1
	TN5	J3	769	Grey-green; Fig. 2.12	TN5/TN6 = sharp and erosive; locally Sub2 occurs and strongly bioturbated	Eol1, Eol2, Eol3, Sub2	Sand	AP1

	TN6	J3	46	Grey; Fig. 2.12	TN6/TN7 = sharp and non-erosive	SFI2, ASt7, ASt6	Sand, pebbles	AP2-AP1
	TN7	K	930	Grey, green, blue, red		SFI1, Wav1, AGr3	Sand, silt	L2, nearly D1
Wusu	W1	J2	65	Grey, dark grey; Fig. 2.13	W1/W2 = sharp but not erosive	AGr2, ASt7, AGr1, ASt4 + OM (locally)	Pebbles, sand	AF3-AF2 Locally D or L
	W2	J2	33	Grey	W2/W3 = sharp but not erosive	ASt6/ASt7, Sub2, ASt2	Pebbles, sand, silt	AP3
	W3	J2	90	Grey	Many lenticular bodies; winnowed surfaces; W3/W4 = sharp but not erosive	ASt3, ASt6, ASt4	Sand, pebbles	AF3-AF2
	W4	J2	55	Grey, brown	Cone-in-cone, stromatolites, plants remains; W4/W5 = non-erosive	Sub1	Silt, sand	Most proximal part of L1
	W5	J2	71	Grey, brown	Bioturbation; W5/W6 = winnowed surface	SGr3, ASt3, Sub2 + OM	Sand, silt	D2 (locally on the edge to the D3 or the D1)
	W6	J2	83	Grey, brown, yellow	Locally coal fragments (cm-long) in sands; W6/W7 = not erosive	ASt6, ASt3	Pebbles to cobbles, sand	AP3-AP2
	W7	J2	60	Grey	W7/W8 = sharp, but not erosive	ASt6, ASt2 + OM	Pebbles to cobbles, sand	AF3-AF2, and some AP
	W8	J2	53	Grey, brown, white, black; Fig. 2.13	W8/W9 = sharp, but not erosive	ASt4, Sub2, ASt6 Tuf1 + OM	Sand, silt	Most distal part of AP3
	W9	J2	64	Grey, brown, white	W9/W10 = sharp, but not erosive	AGr1, ASt7, AGr2 Tuf1	Pebbles to cobbles	AF3 (sometimes closer to the AF2)
	W10	J2	54	Red, grey; Fig. 2.13		ASt6/ASt7, ASt4	Pebbles to cobbles, sand	AF1
	W11	J2	51	Red, grey	Poorly exposed; W11/W12 = sharp, but not erosive	Sub2	Silt	AP1
	W12	J2	118	Grey,	W12/W13 =	ASt2, ASt6	Sand,	AP2

				yellow, white	sharp, but not erosive	Tuf1	conglomerates (pebbles)	
	W13	J2	148	Brown, yellow, grey	W13/W14 = sharp, but not erosive	ASt2, ASt6, Sub2 Tuf1	Sand, conglomerates, silt	AP2-AP1
	W14	J3	37	Grey, red, purple		ASt2, ASt6 Tuf1 + Ped1	Pebbles, sand	AP2
	WE1	J3	14	Red, grey, brown	WE1/WE2 = very erosive	AGr2, ASt6 Tuf1	Silt, sand, pebbles to cobbles	AP3
	WE2	J3	113	Brown, red	WE1/WE2 = sharp boundary	ASt6, ASt2, AGr2 + Ped1, Ped2	Pebbles to cobbles, sand	AP2
	WE3	J3	190	Brown, orange	WE3/WE4 = sharp, but not erosive	ASt2, ASt6 + Ped1, Ped2	Sand, pebbles	AP2, slightly more distal than unit WE2
	WE4	J3	21	Red	WE4/WE5 = sharp and erosive	ASt6, ASt2	Pebbles, sand	In-between AF1 and AP3
	WE5	J3	119	Red; Fig. 2.13	WE5/WE6 = boundary is unclear	AGr2 + ASt7	Conglomerates (pebbles) with sandy matrix	AF2-AF3
	WE6	K/T?	26	Red	Locally some vertical burrows and root traces	ASt2	Silt, sand, granules, pebbles	AP2
Nileke	N1	J2?	299	Grey, brown, black, purple; Fig. 2.14	N1/N2 = sharp boundary	Sub1, ASt2 + OM	Silt, sand	D1 to D2
	N2	J2?	150	Grey, brown, black; Fig. 2.14	Numerous ferruginous concretions	Sub2, ASt4, ASt2 + OM + Ped1	Silt, sand	AP1 and locally L
	N3	J2?	224	Grey, brown, black		Sub2, ASt2 + OM	Silt, sand, pebbles	D3 to D2
	N4	J2?	51	Grey, brown; Fig. 2.14		ASt2	Silt, sand	AP1
	N5	J2?	35	Grey, beige; Fig. 2.14		ASt2, AGr2	Sand, pebbles	AP2 to slightly AP3
	N6	J2?	136	Grey, beige		ASt3, Sub2, AGr2	Pebbles, sand, silt	AP2
Bayanbulak	B1	J2?	42	Grey, black; Fig. 2.15	B1/B2 = sharp	ASt2, ASt6, Sub2 + OM	Pebbles to cobbles, sand, silt	AP3 (slightly more distal than unit

								B2)
	B2	J2?	45	Grey, black	B2/B3 = sharp	ASt2, ASt6, Sub2 + OM	Pebbles to cobbles, sand, silt	AP3
	B3	J2?	64	Grey, beige		ASt7, ASt6 + OM	Pebbles to cobbles	AF3-AF2, locally AP
	B4	J2?	15	Grey, beige; Fig. 2.15		ASt2 + OM	Sand, silt	Most distal part of AP2
	B5	J2?	45	Grey, beige		ASt7	Pebbles to cobbles	AF3
	B6	J2?	72	Grey, beige	B6/B7 = relatively sharp	ASt6 /ASt7, ASt2	Pebbles to cobbles, sand, silt	Proximal part of AF2, with AP3 interbedded
	B7	J2?	60	Grey, beige	B7/B8 = relatively sharp	ASt6 /ASt7	Pebbles to cobbles	AF2-AF3
	B8	Cen.	29	Orange; Fig. 2.15	B8/B9 = relatively sharp	ASt7 Ped3	Pebbles to cobbles, sand	AF3
	B9	Cen.	48	Orange; Fig. 2.15	Very poorly exposed	SFl1 + Ped2	Clayey silt- and sand	Most distal part of AP1
	B10	Cen.	155	Red ; Fig. 2.15		SFl1, AGr2 + Ped1, Ped2	Silt to fine sand, granules, pebbles	AP2
	B11	Cen.	33	Red	B11/B12 = clear boundary	ASt3 + Ped2	Sand	AF2-AF3, locally AP
	B12	Cen.	>24	Red		AGr2, ASt2	granules, pebbles, sand	AP3-AP2
Yaha	Y1	J1	92	White ; Fig. 2.16.1	Y1/Y2 = relatively sharp	SGr3, SGr2, Sub2 + OM	Sand, silt, pebbles	D2
	Y2	J2	112	Grey, beige, black ; Fig. 2.16.1		ASt2 + OM + Ped1	Silt, sand	AP1
	Y3	J2	208	Grey, beige, black	Y3/Y4 = major fault zone, though the displacement is expected to be insignificant for our study	ASt2, SFl1, ASt4, Wav1 (few) + OM	Silt, sand, pebbles	AP2
	Y4	J2	526	Grey, beige, black ; Fig. 2.16.1	Horizontal burrows (cm) ; Y4/Y5 = sharp boundary	*ASt2, SGr3, AFl1, ASt1, ASt3 + OM *SFl2, SGr2	Silt, sand	*D3, overlaid by *D2

Y5	J2	202	Grey, green		SFl1, AS2, SGr2, Cal1 + OM	Sand, silt	L2/L1
Y6	J2	90	Green, grey-blue		SFl1, SGr1, Cal1	Silt, sand	Most distal part of L2
Y7	J3	168	Red (locally green or blue); Fig. 2.16.1	Undetermined gastropods	SFl1, AS2 + Ped2	Silt	D3
Y8	J3	13	Red; Fig. 2.16.1		AS2	Sand, silt	AP2
Y9	J3	30-50	Red ; Fig. 2.16.1	Y9/Y10 = sharp boundary	AS8, AGr2 + Ped2	Pebbles, sand	From AF1 to AF3
Y10	K1	17	Red	Rare horizontal and vertical burrows	AS4, AS2, Sub2 + Ped1	Sand, silt	AP2
Y11	K1	292	Red ; Fig. 2.16.1		Sub2, AS2, AS4, {AS1 reworked by Wav1}, AF1, SGr2 + Ped1, Ped2	Silt, sand	D3
Y12	K1	174	Red; Fig. 2.16.2	Vertical burrows	AS2, Sub2/SFl1, AF1, Wav1, SGr1	Silt, sand	D2 and D3
Y13	K1	366	Red; Fig. 2.16.2	Burrows	AS2, AS4 + Ped1	Silt, sand	AP1
Y14	K1	185	Red-orange; Fig. 2.16.2	Y14/Y15 = very sharp	AS2, AS3, AS4, AF1	Sand, silt	AP1 to AP2
Y15	K2	72	Orange ; Fig. 2.16.2		AGr2, AS7	Pebbles to cobbles	AF2 with some more proximal trends
Y16	K2	169	Orange, purple, white ; Fig. 2.16.2	Y16/Y17 = clear erosive boundary	AS4, AS6 + Ped2	Sand, pebbles	AP2
Y17	K2	6,5	Orange, purple, white ; Fig. 2.16.2		AS6	Pebbles to cobbles, sand	AF3 to AF2
Y18	K2	13	Orange,	Some vertical	AS3, AS6	Sand, pebbles	AP2

				purple, white ; Fig. 2.16.2	burrows	+ Ped2, Ped3, Ped1 (faint)		
	Y19	K2	13	Orange, purple, white ; Fig. 2.16.2		ASt6 + Ped3	Pebbles to cobbles, sand	AF3 to AF2
	Y20	K2	27	Orange, purple, white ; Fig. 2.16.2		Sub2 + Ped3	Silt, sand	AP2

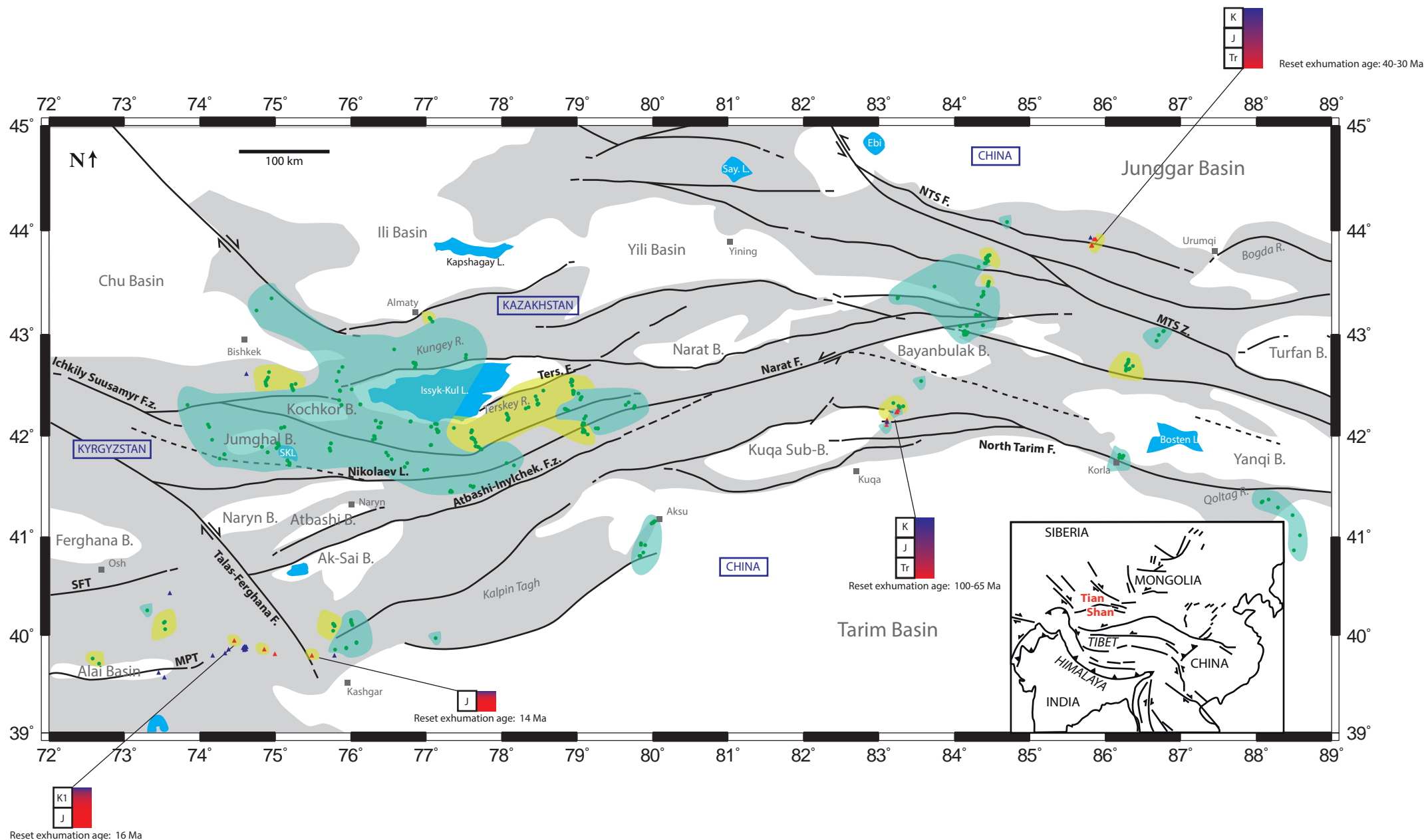
CHAPITRE 3 – SYNTHÈSE DES DONNÉES DE THERMOCHRONOLOGIE BASSE-TEMPÉRATURE (TRACES DE FISSION SUR APATITE, AFT) DANS LA CHAÎNE DU TIAN SHAN ET SES ENVIRONS

Pour reconstruire le relief Mésozoïque dans le Tian Shan et en complément de notre étude sédimentologique (cf. Chapitre 2), il est essentiel de contraindre les principales phases d'exhumation dans la région. Plusieurs auteurs se sont attelés à cette tâche en utilisant des méthodes de thermochronologie basse température. Pour enregistrer les mouvements verticaux dans les deux à trois premiers kilomètres de la croûte continentale, la méthode des traces de fission sur apatite est la plus appropriée. Elle enregistre en effet l'évolution thermique des échantillons entre 120°C et 60°C (ce qui correspond à la zone d'effacement partiel des traces - PAZ, *Partial Annealing Zone*, e.g. Gallagher et al., 1995). En Asie Centrale, la dénudation Cénozoïque est insuffisante pour effacer l'histoire Mésozoïque (cf. Chapitre 1) et peut donc être estimée inférieure à deux ou trois kilomètres. Les traces de fission sur apatite ont par conséquent largement été utilisées pour reconstituer la topographie de cette époque (e.g. De Grave et al., 2011 ; Jolivet et al., 2010). En identifiant les principales phases d'exhumation à l'échelle de la chaîne, il est possible de les comparer à l'histoire enregistrée dans les dépôts sédimentaires. Ces derniers permettent de mieux nuancer et discuter l'intensité des principales phases d'activité tectonique, ainsi que leurs implications en termes de topographie.

Ce chapitre présente une synthèse des données disponibles dans la partie chinoise du Tian Shan, ainsi qu'au Kirghizstan et au Kazakhstan. Trois types de données ont été représentés sur la carte de synthèse. La majeure partie des résultats a été obtenue sur des roches issues du socle (points verts), et ont été regroupées selon l'âge central connu pour chaque échantillon. Dans cette étude, les roches sédimentaires anté-permiennes sont considérées comme du socle, car elles ont été impliquées dans la phase orogénique majeure du Permien et ont de manières générales été remises à zéro. Les deux autres types de données présentées ici ont été acquis sur des échantillons de roches sédimentaires (non-métarmorphisées). Ces dernières peuvent être classées en trois catégories. Soit les sédiments ont été suffisamment enfouis (>120°C) pour que toutes les traces originales soient effacées (*reset* ou remise à zéro) avant la préservation de nouvelles traces lors de l'exhumation et donc du refroidissement de la roche (triangles rouges, Fig. 3.1 et Fig. 3.2 – cas (1)). Les âges individuels des grains (déterminés à partir des traces de fission) forment alors une seule population peu dispersée en général, dont l'âge moyen est inférieur à l'âge de dépôt de la roche. Soit l'échantillon n'a pas été enfoui à plus de 60°C et les grains d'apatite portent encore la signature de

l'exhumation de la roche source, dont ils proviennent (triangles bleus, Fig. 3.1 et Fig. 3.2 – cas (3)). Dans ce cas, la présence d'une ou plusieurs populations d'âges est attendue, toutes antérieures ou contemporaines à l'âge de dépôt (Fig. 3.2). La troisième catégorie rassemble les échantillons intermédiaires, dans lesquels le système a partiellement été remis à zéro (Fig. 3.2 – cas (2)). Certaines populations de grains seront plus anciennes et d'autres plus jeunes que l'âge de dépôt de la roche sédimentaire, ce qui rend l'interprétation de ce type d'échantillon difficile. Notons que l'âge déterminé pour chaque grain d'apatite est l'âge minimum de passage de l'isotherme 120°C.

Figure 3.1 : Carte de synthèse des âges Traces de Fission sur Apatite (AFT) disponibles dans la littérature et obtenus sur des échantillons de socle (points en vert) et de roches sédimentaires (triangles), soit remises à zéro (triangles rouges), soit non remises à zéro (triangles bleus). Les âges centraux (moyenne de la distribution logarithmique des âges individuels pondérés par la précision de mesure sur chaque âge ; Galbraith & Laslett, 1993 ; Gallagher, 1995) ont été utilisés pour réaliser cette synthèse, mises à part les données dans la chaîne du Qoltag pour lesquels seuls les *pooled age* (méthode de calcul qui admet que la totalité des grains comptés proviennent d'une même population d'âge) sont disponibles (Zhang et al., 2011b). Les aires de couleur représentent les zones globalement exhumées durant le Mésozoïque (en bleu) ou durant le Cénozoïque (jaune). La projection de coordonnées utilisée pour cette carte est celle de Mercator. B.= bassin ; Sub-B. = sous-bassin ; L. = lac ; R.= chaîne ; F. = faille ; F. z. = zone de faille ; SKL = Lac de Song-Kul ; Say. L. = Lac Sayram ; Ebi = Lac Ebi ; Ters. F = Faille de Terskey ; NTS F. = Faille du Nord Tian Shan ; Nikolaev L. = Ligne de Nikolaev ; SFT = Chevauchement de Ferghana sud ; MPT = Chevauchement principal du Pamir.



SKL = Song-Kul Lake; CKL = Chatyr-Kul Lake; Kara-Kul Lake; TF = Terskey Fault; NL = Nikolaev Line; STSs = South Tian Shan suture (or Atbashi-Inylchek Fault); SFT = South Ferghana Thrust; MPT = Main Pamir Thrust

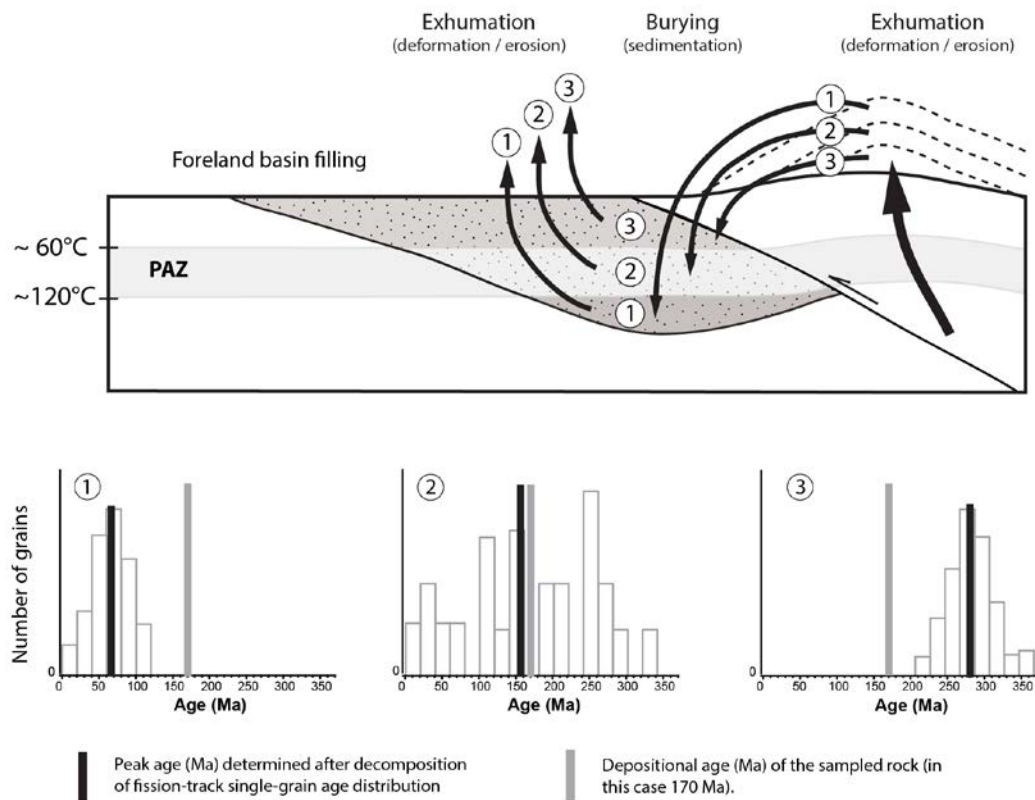


Figure 3.2 : Représentation théorique de l'évolution des âges enregistrés par un grain d'apatite détritique.

Surrection et érosion de la source, dépôt et enfouissement dans le bassin et exhumation des roches sédimentaires. Les échantillons 1, 2 et 3 sont tour à tour érodés, puis déposés dans le bassin. 1 – L'échantillon est enfoui jusqu'à atteindre une température supérieure à 120°C. Toutes les traces de fission sont effacées et le système est remis à zéro. 2 – L'échantillon reste dans la PAZ entre 60 et 120°C et les traces sont partiellement effacées. Le système est partiellement remis à zéro. 3 – L'échantillon n'atteint pas l'isotherme 60°C et ne subit aucun effacement des traces. Le système n'est pas remis à zéro (modifié d'après Braun et al., 2006).

Par conséquent, la présence d'une roche sédimentaire non-remise à zéro indique une zone n'ayant pas été recouverte ultérieurement par d'épais dépôts sédimentaires. Dans le Tian Shan, la disponibilité de matériel à éroder et donc la quantité de sédiments déposés, ont été suffisamment faibles pour ne pas enfouir profondément, ni remettre à zéro les dépôts post-Miocène (cf. dépôts cénozoïques déposés il y a <25 Ma dans le Tian Shan, e.g. Bullen et al., 2001 ; Sobel et al., 2006a ; De Grave et al., 2012). En revanche, durant le Mésozoïque les épaisses séries sédimentaires déposées ont permis la remise à zéro de la partie inférieure de la pile sédimentaires (Echantillon M4 par exemple (Sud-Junggar) Fig. 3.4 – Hendrix et al., 1992 ; voir aussi Fig. 3.1 et Sobel and Dumitru, 1997 ; Dumitru et al., 2001 ; Yang et al., 2014).

Lorsqu'elles sont conservées, les roches sédimentaires mésozoïques non remises à zéro (Fig. 3.1) constituent un indice très précieux pour la reconstitution de la topographie. Elles indiquent une

relative stabilité tectonique (c.à.d. l'absence d'une forte exhumation), qui implique un relatif faible taux d'érosion, qui est lui-même à l'origine d'une faible accumulation de sédiments (maximum 2-3 km d'épaisseur). Il peut donc en résulter la formation d'une surface relativement plane.

Dans la chaîne du Tian Shan, la densité de données est relativement faible compte tenu de l'immensité de la zone ($1 \cdot 10^6 \text{ km}^2$) (Fig. 3.1). Par conséquent, les résultats de cette synthèse ne constituent pas une carte d'exhumation détaillée. Ils ont simplement pour objectif d'identifier et discuter les principales phases d'exhumation, afin de les mettre en parallèle des résultats issus d'autres méthodes (telles que la géomorphologie, la sédimentologie de facies, les analyses de provenances, etc.) pour reconstituer le paléo-relief. Ainsi, des zones regroupant des âges d'exhumation comparables ont pu être identifiées et représentées par des aires de couleurs. Deux périodes d'exhumation dominent : le Mésozoïque (en bleu, 200-65 Ma) et le Cénozoïque (en jaune, <65 Ma) (Fig. 3.1). La phase mésozoïque domine largement les âges d'exhumation. Elle est largement répandue dans les chaînons formant le Tian Shan autour du Bassin d'Issyk-Kul au Kyrgystan (e.g. Kungey au nord et Terskey au sud du bassin). Ces âges mésozoïques sont également conservés au nord de Kashgar et près d'Aksu, dans le sud du Tian Shan. Dans le Tian Shan chinois, en raison des difficultés d'accès les échantillons analysés se concentrent principalement le long d'un transect nord-sud longeant la route qui traverse le Tian Shan chinois (*Du-Ku Road*) et sont également dominés par des âges d'exhumation mésozoïque (Fig. 3.1).

3.1) L'exhumation du socle

Dans les zones étudiées, les roches paléozoïques ou plus anciennes forment le socle de la chaîne. Mis à part les roches formées en profondeur (qui n'ont par conséquent jamais été exhumées), les roches anciennes ont généralement été suffisamment enfouies pour être remises à zéro. Dans la région d'Aksu (Nord-Tarim, Fig. 3.1) par exemple, des roches sédimentaires déposées durant le Protérozoïque et le Paléozoïque ont été enfouies et métamorphisées, avant d'être exhumées (au-dessus de l'isotherme de 120°C) durant le Trias (240-220 Ma) (e.g. Dumitru et al., 2001). Dans les deux parties (ouest et est) de la chaîne, les âges d'exhumation les plus anciens sont triasiques mis à part quelques rares âges permien (Dumitru et al. 2001 ; Sobel et al., 2006a ; Jolivet et al., 2010 ; Macaulay, 2013). Ceci indique que le dernier événement tectonique de grande ampleur a dû avoir lieu au Paléozoïque supérieur (Permien) (cf. Chapitre 1). Localement, ces âges d'exhumation montrent que l'épaisseur des sédiments accumulés depuis le Paléozoïque supérieur

n'a pas été suffisante pour enfouir à nouveau et remettre à zéro ces roches, ce qui a permis la conservation de l'enregistrement de l'exhumation paléozoïque.

3.1.1) Jurassique

Il existe un groupe d'âges d'exhumation vers 200-175 Ma (Jurassique inférieur), obtenus sur des échantillons de socle (Fig. 3.3 c) et e) ; e.g. Dumitru et al., 2001 ; Jolivet et al., 2010 ; De Grave et al., 2011, 2013 ; Macaulay, 2013). En revanche, les âges indiquant une exhumation au Jurassique moyen sont extrêmement rares. Un groupe d'âges Jurassique supérieur allant de 160 Ma à 145 Ma est présent à l'ouest uniquement, tout autour du Bassin d'Issyk-Kul (Fig. 3.3 e) ; Sobel et al., 2006b ; Glorie et al., 2010 ; De Grave et al., 2011, 2013 ; Macaulay, 2013). En comparant ces résultats avec les modèles d'évolution thermique calculés à partir des données de traces de fission, l'exhumation jurassique se résume à l'activité au Jurassique inférieur (Fig. 3.3 e)), liée à l'érosion de la chaîne permo-triasique (e.g. Dumitru et al., 2001 ; Jolivet et al., 2010). La topographie est sans doute toujours significative au début du Jurassique inférieur et s'aplanit au fur et à mesure du Jurassique inférieur à moyen. Ceci est confirmé par la rétrogradation observée dans les milieux de dépôt du Trias supérieur – Jurassique inférieur, comme par exemple dans le Sud-Junggar où les environnements de cône alluviaux laissent place à des plaines alluviales, voire des lacs (Chapitre 2 - Table 2 ; e.g. Hendrix et al., 1992 ; Bian et al., 2010). Au Jurassique moyen, les environnements de dépôt calmes sont observés tout autour de la région du Tian Shan, et montrent localement l'érosion du relief dans les zones plus proximales (cf. Coupe de Wusu ; Fig. 2.17 ; Fig. 2.19.a). Les dépôts de charbon présents dans les bassins du Junggar, du Tarim et de Turfan au Jurassique inférieur et moyen (e.g. Hendrix et al., 1992 ; Shao et al., 2003 ; Jiang et al., 2008 ; Bian et al., 2010) confirment une topographie plane autour du paléo-relief principal à cette époque.

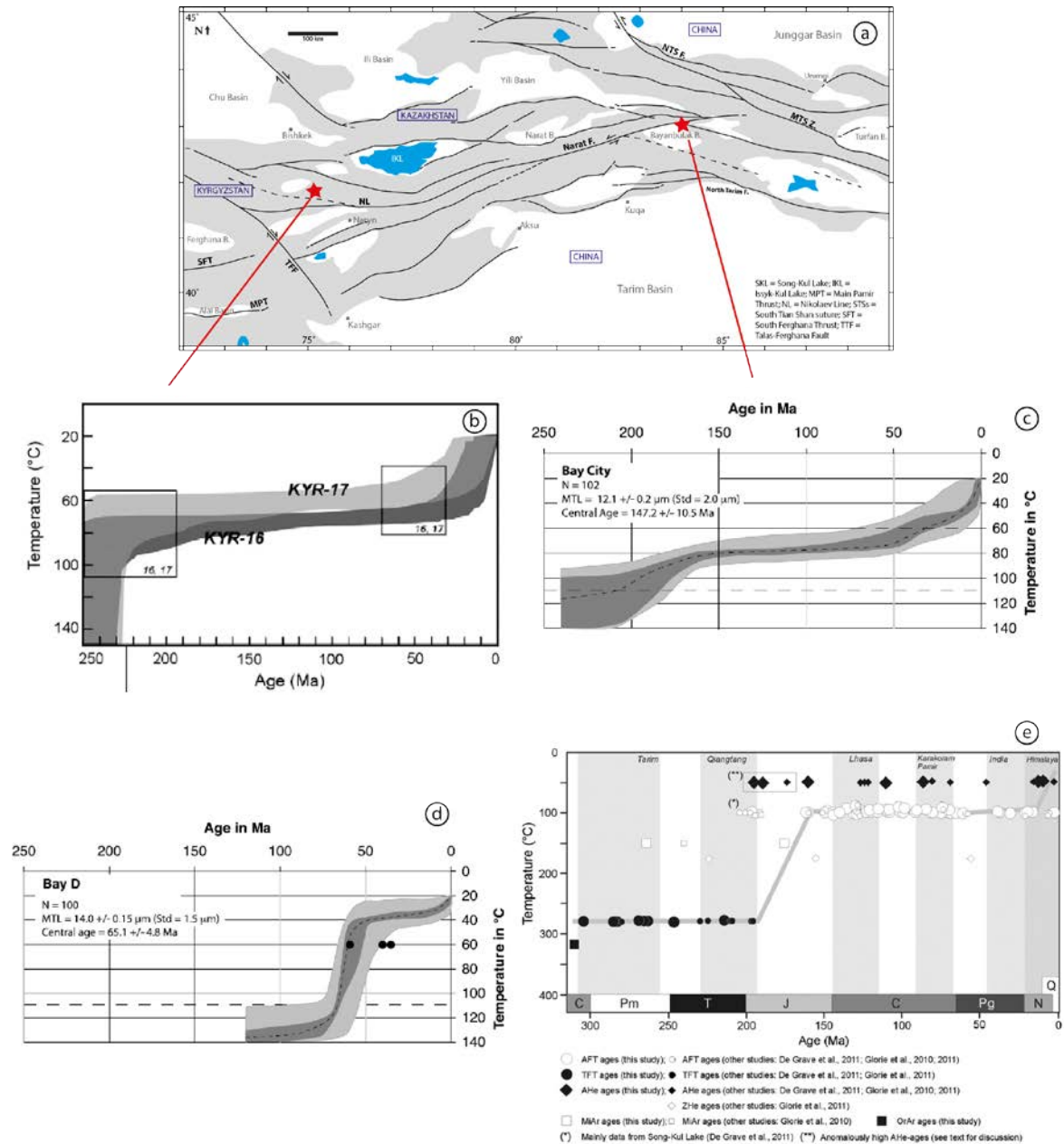


Figure 3.3 : La période de calme tectonique au Jurassique et la réactivation tectonique à la limite Crétacé-Tertiaire. a) Carte de localisation des échantillons dont le modèle est représenté en b), c) et d).

b), c) et d) : Modèles d'évolution thermique calculés à partir des données de traces de fission d'échantillons variés. b) Echantillon situé dans le Tian Shan ouest, au sud du Lac de Song-Kul. Le modèle suggère une phase de stabilité thermique au Jurassique (De Grave et al., 2011). c) Echantillon situé dans le Tian Shan est, au nord du Bassin de Bayanbulak. Le modèle suggère également une phase de stabilité thermique au Jurassique (Jolivet et al., 2010). d) Echantillon situé dans le Tian Shan est, au nord du Bassin de Bayanbulak et à proximité de la Faille de Narat. Le modèle suggère une phase d'exhumation très rapide vers 65 Ma (la limite Crétacé-Tertiaire) (Jolivet et al., 2010). e) Compilation de données de thermochronologie basse-température obtenues à partir d'un grand nombre de données de la région d'Issyk-Kul (Kyrgystan) et en utilisant de nombreuses méthodes d'analyses différentes. Trois grandes phases d'activité tectonique apparaissent : la première au Permien (associée à la collision du Tarim au sud du Tian Shan), la seconde au Mésozoïque (accrétion des blocs de Qiangtang, Lhasa, Karakoram-« Pamir ») et enfin au Cénozoïque la collision Inde-Asie (De Grave et al., 2013).

3.1.2) Crétacé

Au Crétacé inférieur, l'exhumation semble se poursuivre de manière continue à l'ouest (Sobel et al., 2006b ; Glorie et al., 2010 ; De Grave et al., 2011, 2012, 2013 ; Macaulay, 2013), alors qu'il n'existe que très peu d'âges à l'est (e.g. Dumitru et al., 2001 ; Sobel et al., 2006a ; Wang et al., 2009 ; Jolivet et al., 2010). Dans le Tian Shan chinois, cette période est dominée par le dépôt de sédiments lacustres, associés à la présence d'une topographie relativement plane (Fig. 2.18). Il serait intéressant d'étudier plus en détails les sédiments déposés dans le Tian Shan ouest durant le Crétacé inférieur.

En revanche, à la limite Crétacé inférieur – supérieur de nombreux âges AFT (110-100 Ma) semble indiquer une nouvelle phase d'exhumation à l'ouest (Bullen et al., 2003 ; Sobel et al., 2006b ; Glorie et al., 2010, 2012b ; De Grave et al., 2013 ; Macaulay, 2013), comme à l'est (e.g. Dumitru et al., 2001 ; Sobel et al., 2006a ; Wang et al., 2009 ; Zhang et al., 2009 ; Jolivet et al., 2010). A la base du Crétacé supérieur, la formation de cônes alluviaux enregistre une phase de dénudation du relief, et possiblement de réactivation dans le Tian Shan chinois (Coupes de la Manas et Yaha dans les piémonts nord et sud du Tian Shan chinois, Fig. 2.18).

Au Crétacé supérieur, il existe un groupe d'âges d'exhumation compris entre 100 et 65 Ma à l'ouest (au Kyrgystan et Kazakhstan, e.g. Glorie et al., 2010, 2012b ; De Grave et al., 2013 ; Macaulay et al., 2013 ; Macaulay, 2013), comme à l'est (en Chine, e.g. Dumitru et al., 2001 ; Wang et al., 2009 ; Zhang et al., 2009 ; Jolivet et al., 2010). Ces âges sont intermédiaires aux deux grandes phases d'exhumation (permo-triasique et cénozoïque) et indiquent une réactivation continue débutée à la fin du Crétacé inférieur. Par conséquent, il est possible d'imaginer un renouvellement du relief.

3.1.3) Cénozoïque

Plusieurs âges d'exhumation à 65-60 Ma sont observés et confirmés par les modèles d'évolution thermique, qui indiquent une exhumation rapide (Fig. 3.3.d). Ainsi dans le Tian Shan est, ces âges ont été enregistrés le long de la faille de Narat (qui correspond au prolongement de la Ligne de Nikolaev), de la faille de Baluntai et dans le Bogda Shan (Fig. 3.1, 3.3 ; Wang et al., 2008 ; Wang et al., 2009 ; Jolivet et al., 2010). Ceci suggère une réactivation localisée le long des failles. Dans le Tian Shan ouest, des âges équivalents (70-60 Ma) sont observés le long de la zone de failles de Karakunug–Zaili (au nord du Lac Issyk-Kul), le long de la faille de Terskey et de la Ligne de Nikolaev (au sud du Lac Issyk-Kul) (Fig. 3.1, 3.3 ; e.g. De Grave et al., 2013 ; Macaulay, 2013). Ceci indique une réactivation affectant l'ensemble de la chaîne du Tian Shan à la limite Crétacé – Tertiaire (ca. 65 Ma), mais localisée le long des failles. Elle est possiblement due à des répercussions de longues distances,

causées soit par l'accrétion de blocs tels que le Kohistan ou l'arc du Dras, soit par l'extension de la lithosphère dans la zone Sibérie-Mongolie-Chine du Nord (e.g. Jolivet et al., 2010).

Dans les échantillons issus du socle de l'ensemble de la chaîne, les âges cénozoïques montrent une nouvelle phase d'exhumation à partir de 30-25 Ma, qui s'intensifie au Miocène (âges <25 Ma) et possiblement au Pliocène (âge <5 Ma) (e.g. Dumitru et al., 2001 ; Bullen et al., 2003 ; Sobel et al., 2006a, 2006b ; Wang et al., 2009 ; De Grave et al., 2012, 2013 ; Macaulay et al., 2013 ; Macaulay, 2013). Sur la coupe de Yaha (piémont sud du Tian Shan Chinois), les résultats de magnétostratigraphie indiquent une augmentation du taux de sédimentation à partir de 11 Ma (Charreau et al., 2006). Les âges cénozoïques qui précèdent cette phase (entre 65 et 30 Ma) ont subi une exhumation plus lente que les échantillons exhumés au Miocène, et plus lente que ceux exhumés au Mésozoïque. Par conséquent, ces âges intermédiaires peuvent correspondre à une remise à zéro partielle et une exhumation continue (Fig. 3.3.e).

Certains modèles thermiques déduits des données de traces de fission indiquent un refroidissement faible voire nul et donc une période de faible érosion durant le Mésozoïque, ce qui est compatible avec une phase de calme tectonique. La préservation des âges mésozoïques confirme un faible taux d'exhumation durant cette période et indique que la couverture sédimentaire n'a pu excéder 2000 m d'épaisseur (en considérant un gradient géothermique de 30°C/km). En revanche, les séries sédimentaires enregistrant une phase de dénudation significative à la fin du Jurassique (Formation de la Kalaza), elles indiquent bien la présence d'un relief, mais qui ne dépasserait pas deux à trois kilomètres, puisqu'il n'est pas enregistré par la thermochronologie basse température.

De plus, l'érosion cénozoïque a dû être très limitée pour conserver les âges mésozoïques, en dépit de la déformation cénozoïque significative observée à l'ouest (au Kirgystan) et à l'est (en Chine) (e.g. Avouac et al., 1993a ; Burbank et al., 1999 ; Charreau et al., 2008 ; Glorie et al., 2011a, 2012b ; De Grave et al., 2013).

Les zones où sont conservés les âges d'exhumation mésozoïques sont généralement éloignées des failles actives et correspondent aujourd'hui à des plateaux (localement des surfaces planes basculées), dont les bordures indiquent des âges d'exhumation cénozoïques. Ces surfaces planes ont été observées au Kirgystan au sud du Bassin de Naryn (e.g. Burbank et al., 1999 ; Glorie et al., 2010) par exemple. Elles sont recouvertes de dépôts cénozoïques discordants, et le tout a été basculé en conséquence des réactivations et déformations cénozoïques de la chaîne (Fig. 3.4). Allen

et al. (2001) mentionnent également la présence d'une pénéplaine Crétacé-Paléogène dans la zone de Karatau (sud du Kazakhstan). Les surfaces de pénéplaines sont également présentes à l'est dans le Tian Shan chinois et en Mongolie (e.g. Jolivet et al., 2007, 2010).



Figure 3.4 : Photo illustrant la surface de pénéplaine, recouverte de dépôts paléogènes (orange), le tout étant basculé. Détails dans Cobbold et al. (1994) et Burbank et al. (1999) ; photo M. Jolivet (2012).

En revanche, les âges d'exhumation cénozoïques indiquent une remise à zéro du système. Traces de fission dans le socle et une nouvelle phase d'exhumation. Les zones exhumées au cours du Cénozoïque sont moins vastes et moins répandues que celles du Mésozoïque. Toutefois, l'exhumation semble être plus prononcée à l'ouest qu'à l'est. Dans le Tian Shan chinois, les âges cénozoïques semblent être majoritairement localisés dans les piémonts (nord et sud) de la chaîne, ainsi que le long des principales zones de failles actives.

Certaines zones comportent des âges d'exhumation mixtes, avec des âges mésozoïques et cénozoïques. En général, ceci correspond à un profil vertical échantillonné au travers des chaînons exhumés durant le Cénozoïque. Lors de réactivations tectoniques les zones de failles pré-existantes

représentent des zones de faiblesse qui sont les premières à être réactivées et donc exhumées (e.g. Burbank et al., 1999 ; Jolivet et al., 2010 ; Glorie et al., 2012b).

3.2) L'exhumation des roches sédimentaires

Dans les échantillons de roches détritiques plusieurs populations de grains (ayant des âges d'exhumation différents) peuvent être identifiées et reflètent les principales phases de dénudation, et par conséquent d'exhumation. Les échantillons analysés proviennent des bassins kirghizes à l'ouest, et des piémonts nord et sud de la chaîne à l'est.

Ces échantillons servent par exemple à déterminer à partir de quel niveau stratigraphique l'enfouissement a été suffisant pour les remettre à zéro, du moins partiellement. Par exemple, dans le piémont nord du Tian Shan chinois, cinq échantillons sédimentaires provenant d'une même section ont été analysés. Les trois plus anciens stratigraphiquement (d'âge triasique et jurassique) ont été exhumés vers 40-30 Ma (Echantillon M4, Fig. 3.5, Dumitru et al., 2001). Dans le quatrième échantillon, déposé à la fin du Jurassique, la répartition des âges individuels montre que certains grains d'apatite indique un âge d'exhumation antérieur au dépôt et d'autres grains un âge postérieur. Ceci indique une remise à zéro partielle. Enfin, dans le dernier échantillon d'âge crétacé, les âges des apatites ne forment qu'une seule population d'âge jurassique (187 Ma) et indique que l'échantillon n'a pas été remis à zéro (Dumitru et al., 2001). Ceci indique également que durant le Crétacé, ce sont principalement les roches exhumées au Jurassique inférieur qui ont été érodées. Ces résultats sont en accord avec les âges d'exhumation au Jurassique inférieur obtenus sur les échantillons de socle (cf. 3.1 Exhumation du socle). Grâce aux AFT, il est donc possible de retrouver des contraintes sur les sources des sédiments, même si la roche source a été érodée et a donc disparu.

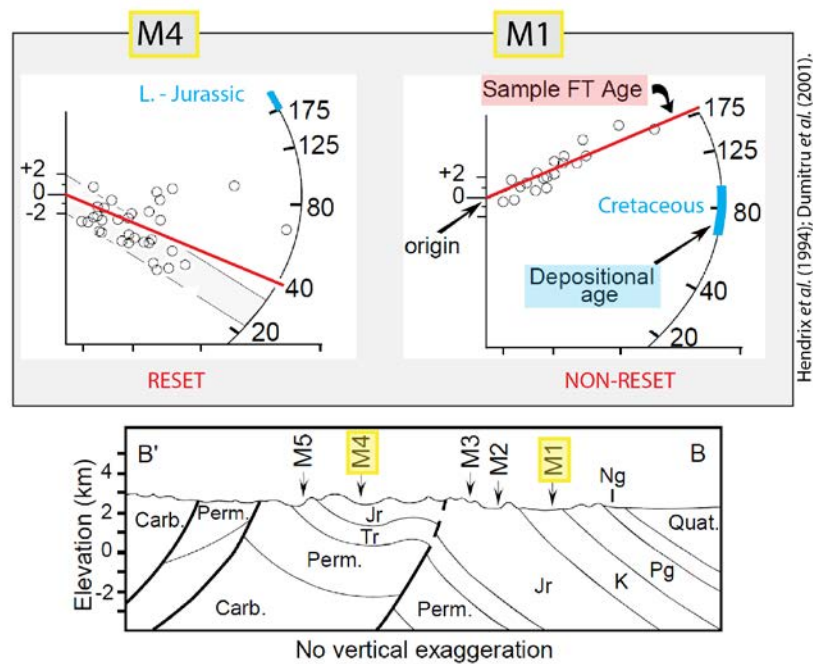


Figure 3.5 : Assemblage de figures extraites de Hendrix et al., 1994 et Dumitru et al., 2001), montrant le profil d'échantillonnage (en bas) de la série de sédiment mésozoïques (Coupe de la Manas) et les diagrammes radiaux de deux des échantillons. Les échantillons de roches sédimentaires M1 et M4 sont issues de la coupe de la Manas (Sud-Junggar, à l'ouest d'Urumqi ; Fig. 3.1). Les diagrammes radiaux (*radial plot*) montrent la répartition des âges individuels des grains détritiques. Chaque cercle blanc indique l'âge d'un grain d'apatite. L'âge du dépôt de la roche sédimentaire est indiqué en bleu pour comparaison. L'échantillon M4 déposé au Jurassique inférieur ne comporte que des grains d'âge postérieur au dépôt. Par conséquent, cet échantillon a été remis à zéro. En revanche, les grains individuels présents dans l'échantillon M1 (déposé au Crétacé) sont tous antérieurs à l'âge de dépôt et l'échantillon n'est pas remis à zéro. La coupe schématique indique la continuité des séries sédimentaires Méso-Cénozoïques. L'échantillon M1 n'étant pas remis à zéro, il est possible d'en déduire que les sédiments cénozoïques ne le sont pas non plus.

Les dépôts sédimentaires jurassiques et crétacés non remis à zéro contiennent quasi-exclusivement des grains aux âges d'exhumation très anciens, allant du Permien supérieur au Jurassique inférieur (270 – 187 Ma), et ceci dans l'ensemble de la chaîne du Tian Shan (à l'ouest comme à l'est) (Fig. 3.4; Dumitru et al., 2001 ; De Grave et al., 2012).

En revanche, dans les échantillons déposés au cours du Cénozoïque, plusieurs populations d'âges apparaissent. Au Kirgystan, les dépôts néogènes de la région de l'Alai (De Grave et al., 2012), et les dépôts miocènes-néogènes à l'ouest du Lac Issyk-Kul (<10 Ma) (Bullen et al., 2001) comportent des âges d'exhumation anciens : Permien supérieur ou Triasique (223-205 Ma dans l'Alai et 260 – 230 Ma près du Lac Issyk-Kul). Les âges les plus jeunes sont inférieurs à 25 Ma dans les deux zones et représentent une des principales populations. Entre ces deux populations d'âges extrêmes, des âges jurassiques sont également présents et représentent une autre large population. Il existe quelques âges d'exhumation crétacés supérieurs (80-70 Ma).

Dans l'extrême nord-ouest du Bassin du Tarim, c.à.d. à une centaine de kilomètres à l'est des échantillons de la région de l'Alai (Fig. 3.1) analysés par De Grave et al. (2012), Yang et al. (2014) ont également obtenus des âges d'exhumation dans les sédiments. Un des échantillons déposés à l'Eocène-Oligocène ne comporte aucun âge supérieur à 100 Ma et est dominé par les âges <60 Ma. Dans les dépôts miocènes qui suivent, les âges maximum sont compris entre 110 et 75 Ma et les plus jeunes vers 26 Ma. Cette absence des populations d'âges anciens n'est pas expliquée, mais pourrait être la conséquence du biais que représente la difficulté d'analyser les grains les plus anciens et souvent les plus abimés.

Sur la coupe de Wenguri (une centaine de kilomètres plus à l'est d'Ulugqat, Fig. 3.1), Sobel & Dumitru (1997) obtiennent des âges d'exhumation répartis en trois populations : 145-100 Ma, 50-35 Ma et <25 Ma.

De manière générale, la synthèse de ces données montre que dans les sédiments cénozoïques, les âges relativement anciens (pré-Crétacé) disparaissent pour laisser place aux nombreux âges cénozoïques. Les échantillons de sédiments déposés depuis le Miocène (< 23 Ma) sont dominés par des âges d'exhumation inférieurs à 25 Ma. Ceci confirme la réactivation de l'exhumation à partir du Miocène (e.g. Sobel & Dumitru, 1997 ; Bullen et al., 2001 ; Dumitru et al., 2001 ; De Grave et al., 2012), également indiqué par les analyses AFT sur des roches issues du socle de la chaîne. De plus, ceci montre que le socle a été exhumé et érodé, et le produit de cette érosion a été déposé très peu de temps après, indiquant une exhumation très rapide de la zone source. Dans la plupart des échantillons néogènes, l'âge AFT le plus jeune des populations détritiques est très proche de l'âge de dépôt.

A l'est, on observe également la perte des âges très vieux (>150 Ma, pré-Crétacé) pour laisser place aux âges jeunes (<50 Ma) dans les dépôts de la fin du Miocène (vers 5 Ma sur la coupe de la rivière Kuitun, piémont nord du Tian Shan chinois) (cf. Annexe).

A partir du Pliocène et au Quaternaire, deux échantillons analysés près d'Ulgqat (Fig. 3.1) montrent une baisse de la proportion d'âges jeunes, accompagnée de la présence d'âges plus anciens que précédemment (130-30 Ma). Ceci peut être interprété soit comme une diminution de l'exhumation ou comme un changement dans les sources (Yang et al., 2013). Dans le Tian Shan chinois, les âges très jeunes (<10 Ma) semblent tout de même être plus rares que dans l'ouest de la chaîne (cf. Annexe). Ceci confirme la dichotomie entre les deux parties de la chaîne et laisse à penser que l'exhumation semble moins prononcée à l'est.

Cette différence est également observable dans l'âge d'exhumation des échantillons détritiques remis à zéro. Les échantillons déposés au Jurassique et Crétacé dans le nord-ouest du Bassin du Tarim ont été exhumés vers 16-14 Ma d'années et sont des échantillons remis à zéro (Sobel & Dumitru, 1997 ; Yang et al., 2014). Tandis que les échantillons jurassiques et crétacés des piémonts nord et sud plus à l'est n'ont pas tous été remis à zéro. La partie est du Tian Shan semblerait être affectée par une exhumation moins prononcée qu'à l'ouest, mais débutant plus tôt (100-65 Ma sur la coupe au nord de Kuqa et 40-30 Ma sur la coupe de Manas, Fig. 3.1).

3.3) Interprétation à l'échelle de la chaîne

La synthèse des résultats de thermochronologie (méthode AFT) montre que l'ensemble de la chaîne n'a pas subi d'exhumation significative au cours du Mésozoïque supérieur (Jurassique moyen à Crétacé), ce qui complète notre analyse sédimentologique (Chapitre 2). Pourtant au Jurassique supérieur, la présence de cônes alluviaux dans les piémonts indique la présence de reliefs, qui ne sont pas enregistrés par les traces de fission. Nous suggérons que ces reliefs n'aient simplement pas généré une exhumation suffisante (soit deux à trois kilomètres) pour être enregistrés par le système traces de fission. Ainsi, la combinaison des deux méthodes a permis de résoudre l'apparent paradoxe du Jurassique supérieur.

A l'ouest, les sédiments mésozoïques sont moins bien exposés et conservés que dans le Tian Shan chinois. Le Tian Shan chinois est donc la seule région offrant de bonnes conditions d'affleurement des séries mésozoïques. La synthèse des données AFT indique une dichotomie entre les zones ouest et est. Il semblerait que l'exhumation cénozoïque ait plus affecté la partie ouest du Tian Shan. Cette différence peut être expliquée par la rotation du bloc du Tarim dans le sens horaire, et dont le centre de rotation (cénozoïque) serait situé dans l'est du Tarim (e.g. Avouac et al., 1993a). Cette rotation implique un raccourcissement plus fort à l'ouest qu'à l'est. Il en résulte que l'exhumation différentielle est forte entre les deux zones (est et ouest).

En termes de paléo-relief, la faible quantité d'exhumation mésozoïque semble confirmer une chaîne de faible altitude (ou une topographie relativement plate) durant le Jurassique et le Crétacé. De plus, la conservation des surfaces mésozoïques montre que la quantité d'exhumation cénozoïque a été relativement restreinte jusqu'à présent. Ceci est en accord avec la formation d'une vaste

pénéplaine dans toute l'Asie centrale et sa préservation jusqu'à aujourd'hui (e.g. Dumitru et al., 2001 ; Jolivet et al., 2001 ; Vassallo et al., 2007b ; Jolivet et al., 2007, 2010 ; Glorie et al., 2010).

Un régime tectonique transtensif ou extensif comme suggéré dans le chapitre précédent, semble donc être en cohérence avec ces observations, contrairement à la majorité des études précédentes qui argumentent pour une réactivation de la chaîne et un régime compressif (e.g. Hendrix et al., 1992 ; Dumitru et al., 2001).

Dans l'Altay plus au nord, l'exhumation a lieu à deux moments : au Jurassique supérieur – Crétacé inférieur et au Pliocène (8-3 Ma) (e.g. De Grave and Van den haute, 2002 ; De Grave et al., 2008 ; Briggs et al., 2009 ; Glorie et al., 2012a). La première phase d'exhumation semble être associée à une réactivation et peut être mise en parallèle avec la fermeture de l'océan Mongol-Okhotsk (e.g. De Grave and Van den haute, 2002 ; De Grave et al., 2007, 2008).

3.4) En Asie Centrale

Dans le Tian Shan, les âges traces de fission sur apatite indiquent globalement trois principales phases d'exhumation. Une première phase a lieu à la fin du Paléozoïque, suivie d'une phase tardi-mésozoïque et de la phase d'exhumation cénozoïque. La première est généralement associée à la collision entre l'ensemble Tarim – Tian Shan Central et le Tian Shan (limite Carbonifère – Permien). Elle correspond aussi à la période d'amalgamation finale de la CAO B en Asie Centrale. La seconde peut être mise en relation avec l'accrétion du bloc de Lhasa sur la marge sud du continent asiatique (Jurassique - Crétacé). La dernière phase d'exhumation est communément interprétée comme étant la réactivation de la chaîne due à la collision entre l'Inde et l'Asie (Paléogène) (e.g. Dumitru et al., 2001 ; Jolivet et al., 2010 ; De Grave et al., 2013). Généralement les âges sont dominés par la phase d'exhumation mésozoïque, comme c'est également le cas plus à l'est en Mongolie, plus au sud dans le Nord-Tibet ou plus au nord en Sibérie (e.g. Jolivet et al., 2001, 2007, 2009 ; Vassallo et al., 2007).

3.5) Résumé

Les âges d'exhumation (AFT) indiquent trois grands groupes : après une petite phase d'exhumation au Jurassique inférieur, le premier groupe indique un calme apparent durant le Jurassique moyen et supérieur ; le second groupe montre une exhumation continue durant le Crétacé avec une accélération vers 100 Ma surtout à l'ouest et une vers 65 Ma) et enfin le dernier

groupe comprend les âges d'exhumation tertiaires avec une accélération à partir du Miocène (<25 Ma). Les deux principales phases d'exhumation sont donc mésozoïques (aires coloriées en bleu sur la Fig. 3.1) et tertiaire (aires coloriées en jaune sur la Fig. 3.1), la première étant largement dominante.

La conservation de ces âges indique une exhumation relativement faible depuis le Mésozoïque supérieur, ainsi qu'une quantité de dépôts raisonnable (moins de 2-3 km d'épaisseur). Par conséquent ceci implique également que l'exhumation tertiaire qui a suivi, devait être relativement faible pour ne pas effacer l'enregistrement précédent. Ces conditions ont ainsi permis la préservation d'une surface de pénéplanation, qui a également été observée ailleurs en Asie Centrale (Mongolie, etc.).

En complément des analyses sur le socle, les différentes populations d'âges d'exhumation conservées dans certaines roches sédimentaires, confirment la faible exhumation depuis le Mésozoïque supérieur. En plus des variations des sources, elles indiquent également les zones exhumées très rapidement suite à leur dépôt (âge d'exhumation le plus jeune très proche de l'âge de dépôt).

Dans cette synthèse, il apparaît que le Tian Shan kirgize et kazakhe semble avoir subi plus d'exhumation que le Tian Shan chinois, où les roches exhumées au Tertiaire sont surtout localisées le long des zones actives et où les sédiments mésozoïques n'ont pas encore totalement été remis à zéro. Cette dichotomie entre l'ouest et l'est semble être le résultat de la rotation du bloc du Tarim, qui implique un raccourcissement plus fort à l'ouest.

La combinaison des résultats de l'analyse des sédiments (Chapitre 2) et de cette synthèse permet de lever le voile sur l'apparent paradoxe du Jurassique supérieur. Nous proposons que les reliefs suggérés par la présence de cônes alluviaux dans les piémonts, n'aient pas été suffisamment exhumés (soit deux à trois kilomètres) pour être enregistrés par le système traces de fission. Le contexte tectonique général, ainsi que l'ensemble des données (sédimentologiques et thermochronologiques) sont incompatibles avec un système en compression. Nous suggérons la présence d'un système de failles normales, créant des grabens ou demi-grabens dans un contexte extensif au Jurassique supérieur et au Crétacé inférieur. Ceci est cohérent avec le contexte extensif qui prévaut en Asie Centrale à cette époque. La réactivation cénozoïque, quant à elle correspond aux effets de grandes longueurs d'ondes dus à la collision Inde-Asie.

Stable isotope characterization of pedogenic and lacustrine carbonates from the Chinese Tian Shan:
constraints on the Mesozoic – Lower Cenozoic palaeo-environmental evolution

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Abstract

In the Mesozoic-Cenozoic continental deposits of the Tian Shan area, two main levels containing pedogenic carbonates have been identified on both the southern and northern foothills of the range: one in the Upper Jurassic series and one in the Upper Cretaceous - Lower Palaeocene series. In order to reconstruct the palaeo-environmental and palaeo-topographic characteristics of the Tian Shan area during these two periods, we measured the oxygen and carbon isotope composition of these pedogenic carbonates (calcrete and nodules). The stable isotope compositions are homogeneous: most $\delta^{18}\text{O}$ values are between 21 and 25‰ and most $\delta^{13}\text{C}$ values are between -4 and -6‰. No distinction can be made between the calcrete and nodule isotopic compositions. The constancy of isotopic values across the Tian Shan is evidence of a development of these calcification features in similar palaeo-environmental conditions. The main inference is that no significant relief existed in

that area at the Cretaceous - Palaeogene boundary, implying that most of the present relief developed later, during the Cenozoic. In addition to the pedogenic carbonates, few beds of limestones interstratified in the Jurassic series of the southern foothills display oxygen and carbon isotope compositions typical of lacustrine carbonates, ruling out brackish water incursion at that period in the region.

Keywords: pedogenic carbonates - stable isotopes - Mesozoic palaeo-environment - Tian Shan

1. Introduction

Calcification is a ubiquitous phenomenon that occurs in a large variety of geological settings including limestone diagenesis and continental weathering. In continental environments, under semi-arid conditions, the interaction between meteoric waters or groundwaters and sedimentary rocks commonly leads to pedogenic calcification (e.g. Wright and Tucker, 1991; Alonso-Zarza, 2003; Hasiotis et al., 2007). The resulting calcareous features correspond to the so-called calcretes and occur as isolated nodules or rather continuous levels of carbonates (Retallack, 1997; Wright, 2008). The oxygen and carbon isotope composition of such neo-formed carbonates provides information on the palaeo-environments within which they precipitated (e.g. Alonso-Zarza, 2003). Indeed, while the oxygen isotope composition of these carbonates depends upon processes taking place in the soil – ground water system, such as evaporation, it is also function of the composition of the meteoric water. The latter depends on the palaeo-latitude and the palaeo-altitude of the area, as well as on the distance to the ocean from which the atmospheric masses are derived. In a specific region, any variation in the $\delta^{18}\text{O}$ value of the carbonates through time may thus be, for example, a consequence of the climatic evolution. At a specific stratigraphic level, geographic variations of $\delta^{18}\text{O}$ may relate to the existence of a significant palaeo-topography (e.g. Garzzone et al., 2000). In a similar way, the carbon isotope composition of continental neo-formed carbonates depends on the source of the carbon in the region, which relates to the relative abundance of carbon-bearing rocks (e.g. marine carbonates) over available organic carbon. The former have $\delta^{13}\text{C}$ values close to 0‰, whereas the second source has very negative values, close to -25‰ for carbon derived from C3-type plants and -15‰ from C4-type ones. The $\delta^{13}\text{C}$ signal can thus potentially be related to the geology of the area, the palaeo-flora itself partially controlled by the palaeo-climate (e.g. Pustovoytov, 2002).

In North-West China, the intracontinental, east-west oriented Tian Shan Range (Fig. 1) provides spectacular outcrops of Mesozoic-Cenozoic clastic sediment series. During fieldwork, we recognized several occurrences of calcretes and nodules throughout these series, especially in the Upper Mesozoic and Lower Cenozoic deposits. A strong aridification occurred during Late Jurassic, following the progressive appearance of dry seasons since the late Early Jurassic (e.g. Allen et al., 1991; Hendrix et al., 1992; Eberth et al., 2001; Li et al., 2004; Ashraf et al., 2010; Pan et al., 2012). Since this period, arid to semi-arid climate has been predominant in the Tian Shan area (e.g. Allen et al., 1991; Hendrix et al., 1992; Li et al., 2004; Sun and Wang, 2005; Jiang et al., 2008).

Since the late 1970s, the Cenozoic tectonic development of the range and its foreland basins has been widely studied (e.g. Tapponnier and Molnar, 1979; Windley et al., 1990; Allen et al., 1991; Hendrix et al., 1994; Charreau et al., 2005, 2009, 2012; Jolivet et al., 2014). In contrast, the Mesozoic evolution of the Tian Shan area is still largely debated (Jolivet et al., 2013), despite some efforts to reconstruct the sedimentary depositional environments (e.g. Hendrix et al., 1992; Li et al., 2004; Eberth et al., 2001; Vincent and Allen, 2001; Sha et al., 2011), the palaeo-topographic evolution (Dumitru et al., 2001; Jolivet et al., 2010; Yang et al., 2013; D. Liu et al., 2013; Yang et al., n.d. in press) and the palaeo-climate (e.g. Hendrix et al., 1992, 1994; Hendrix, 2000; Eberth et al., 2001; Li et al., 2004; Ashraf et al., 2010; Bian et al., 2010). In this paper, we provide new isotopic data on pedogenic carbonates (calcrete and nodules) and limestone beds in order to improve our understanding of the general evolution of the area from Late Jurassic to Early Palaeogene.

2. Geological setting

The Tian Shan is a large mountain belt extending in Central Asia through Kyrgyzstan, Kazakhstan and North-West China (Fig. 1a). This range is surrounded by several basins among which the Junggar Basin to the north and the Tarim Basin to the south (Fig. 1b). In the central part of the Tian Shan, several intramontane basins are preserved such as the Yili, Bayanbulak and Turfan basins (Fig. 1 and 2). The Tian Shan Range is limited by large crustal thrusts verging northward in the north and southward in the south towards the Junggar and Tarim Basins respectively (Fig. 1b). During the Palaeozoic, several accretion events gave rise to an important topography that was progressively eroded from Middle Triassic to Middle Jurassic (e.g. Dumitru et al., 2001; Li et al., 2004; Jolivet et al., 2010; 2013; Yang et al., 2013; S. Liu et al., 2013). The present-day topography formed later, mostly through the Cenozoic reactivation of Late Palaeozoic tectonic structures, driven by the far-field effects of the India-Asia collision event (e.g. Tapponnier and Molnar, 1979; Windley et al., 1990; Allen

et al., 1991; Hendrix et al., 1994; Glorie et al., 2010; De Grave et al., 2011, 2013). Whereas the present-day topography is believed to have developed since Late Oligocene – Miocene (e.g. Charreau et al., 2005, 2009, 2012), uncertainties remain about the topographic evolution of the range during the Jurassic - Early Palaeogene period. For example, while low-thermochronology data suggest a long tectonic quiescence during most of the Mesozoic leading to very slow and constant exhumation within the range (Dumitru et al., 2001; Jolivet et al., 2010; 2013), Upper Jurassic alluvial fan deposits exposed in the foothills are interpreted as the consequence of a compressive reactivation of the range (Hendrix et al., 1992; Li et al., 2004). In this context, the present study on calcretes and nodules provides independent information on the palaeo-topographic evolution of the area.

3. Field work and samples

3.1 Field description

Throughout the area, we recognized several calcareous features (calcretes and nodules) at distinct stratigraphic levels (Fig. 2). Very few age data are available due to the lack of clear marine deposits, the scarcity of interbedded volcanic levels and the very limited available palynological and faunistic data (e.g. Wang and Gao, 2012; Yang et al., 2013). Nevertheless regional stratigraphic correlations can be made using available geological maps and the stratigraphic columns presented in published palynoflora and vertebrate studies (e.g. Eberth et al., 2001; Bian et al., 2010; Jiang et al., 2008; Heilbronn, 2014), as well as our own observations.

Below is a short description of the stratigraphy and the inferred associated depositional environments (Heilbronn, 2014). The present-day semi-arid climate of this region offers excellent outcrops preservation and exposure. In the northern foothills, from west to east (Fig. 1), the Wusu section exposes 1650 m of Middle Jurassic to Upper Cretaceous series, the Manas section exposes 8100 m of Triassic to Neogene series and the Qiu Er Gou section exposes 55 m of late Upper Cretaceous deposits. In the Bayanbulak intramontane basin (Fig. 1), 620 m of Jurassic to Neogene series are exposed, the stratigraphic continuity of which remains questionable. Finally, in the southern foothills, the Yaha section provides 2800 m of Jurassic to Neogene deposits.

The Mesozoic series of the Tian Shan area are constituted of continental deposits (Hendrix et al., 1992; Li et al., 2004; Jiang et al., 2008; Li and Peng, 2010). These series are mainly composed of stream and alluvial plain deposits (few decimetres- to a few metre-thick grey and red sandstone beds) alternating with floodplain or lake deposits (clays and siltstones). Thick conglomerates which

deposited in an alluvial fan environment are exposed at the end of the Upper Jurassic and the Upper Cretaceous series (Fig. 2).

Calcareous features are mainly recognized at two intervals: the late Middle Jurassic to Lower Cretaceous and the late Upper Cretaceous to Palaeocene. The most continuous level of neo-formed carbonate is located in the later. This calcrete consists of a several metres thick indurated and white layer, parallel to the bedding and laterally continuous over tens of kilometres (Fig. 3a). The calcrete is commonly associated with conglomeratic beds (host rock), consisting in matrix-supported conglomerates with sub-angular to sub-rounded pebbles (Fig. 3b). Carbonate nodules are white, crumbly to highly indurated, spherical to elongated concretions of micritic carbonate (few centimetres in diameter), and either occur as disseminated bodies in red silty to sandy levels or form coalescent beds. Such a feature is classical for pedogenic carbonate (see Fig. 3E in Li et al., 2013). On several locations, one can observe a vertical transition from a zone enriched in nodules towards the massive calcrete layer (Fig. 3c); both features thus likely represent two stages of calcification, complete for calcrete, incipient for nodules. Processes controlling the formation of massive calcretes are still discussed and it is likely that both phreatic and pedogenic processes are involved (e.g. Wright, 2008 and references therein).

Moreover some limestone beds occur in the silty to sandy alternations of the late Middle Jurassic continental series of the Yaha section (Fig. 2) (e.g. Jiang et al., 2008). They are white to dark grey in colour, several centimetres- to several decimetres-thick and rather continuous laterally (Fig. 3d) and often show evidence of bioturbation (burrows). Middle Jurassic brackish water incursions have been inferred along the southern margin of the Junggar Basin (Chen and Zhang, 2000; Sha et al., 2011; Pan et al., 2013). We thus analysed those limestone beds in order to check their possible marine origin, even if they are interbedded with sediment series described as continental deposits (Hendrix et al., 1992; Sobel, 1999; Jiang et al., 2008).

3.2 Sampling strategy

The recognition of nodules and calcrete layers in several places throughout the Tian Shan Range and their localization in the same stratigraphic levels (Upper Jurassic, Upper Cretaceous) offer a unique opportunity to specify the palaeo-topographic evolution of the area during the late Mesozoic. Every observed type of carbonated pedogenic feature has been sampled. Samples were first collected along a north-south transect across the Tian Shan, mostly focusing on the Manas,

Bayanbulak and Yaha sections. For the Upper Cretaceous level, five to six samples have been collected from each section (Table 1; Fig. 2). When the calcrete and nodules were present together in a single site, both were sampled as they might have formed under distinct pedogenic conditions, even if some continuity between the two types of objects has been locally observed (Fig. 3c). For the Jurassic period, the sampling is not fully complete since calcification has not been observed in the Bayanbulak section and is only poorly developed in the Manas and Wusu sections.

On the northern foothills, six samples were also collected from the Qiu Er Gou section only 20 km east of the Manas section (Fig. 1). Calcification in these two close-by sites is located in the Upper Cretaceous series and likely developed under identical climatic and topographic conditions. They thus allow specifying the isotopic heterogeneities solely associated with the calcification processes. For example, evaporation can induce kinetic effects that modify the $\delta^{18}\text{O}$ signature of pedogenic carbonates (Ufnar et al., 2008 and references therein).

Finally, several late Middle Jurassic limestone beds have been sampled on the Yaha section (Fig. 3d) both in order to confirm or infirm their suspected lacustrine origin using the isotopic tool, and to use their isotopic composition as a complementary proxy to palaeo-environmental reconstruction (e.g. Li et al., 2013).

3.3 Sample description and preparation

A total of 32 fist-sized samples have been collected (Fig. 4). Calcrete samples consist of conglomerates in which the matrix has been cemented by neo-formed calcite (Fig. 4a). Figure 4a shows white calcite cement that precipitated around the pebbles. This cement has been sampled by scraping or micro-drilling and reduced to powder for isotopic analysis. Since the conglomerates contain some detrital fragments of carbonate rocks, special care was taken to avoid contamination of the sampled powder by pebble-derived carbonate. In one sample of the Manas section (sample MC 11-3; Table 1), three types of neo-formed carbonates were observed and sampled: calcitic cement (fraction A), a few millimeters wide veinlets (fraction B) and millimetric geodic calcite grains (fraction C).

Nodule samples consist of nearly pure neo-formed calcite, sometimes easily extracted from their sandy to silty host rock (Fig. 4b). These nodules were crushed and neo-formed calcite was separated by hand-picking from remaining small detrital grains. For both the calcrete and nodule samples, the separated calcite was then crushed in a boron carbide mortar before analysis.

Limestone samples sometimes contain some detrital materials, but no pebble was observed. Centimetre-sized fragments of limestones have been detached from the samples, crushed in an Abiche mortar, then in a boron carbide mortar to get a finely grained powder.

4. Analytical techniques

The stable isotope analysis was performed at the stable isotope laboratory of Geosciences Rennes, University of Rennes 1. About 12 mg of calcite powder have been reacted with anhydrous H_3PO_4 for a few hours at 50°C. The liberated CO_2 was analysed on a VG Optima triple collector mass spectrometre. Routine analysis of in-house standard (Prolabo Rennes) and international standards NBS 18 and NBS 19 ensures accuracy of the results. The global precision is estimated at $\pm 0.15\text{‰}$ for oxygen (SMOW) and $\pm 0.10\text{‰}$ for carbon (PDB). Isotopic results are reported using the conventional "delta" notation in Table 1 and summarized in Figure 5, where they are compared with data from the literature.

5. Results

In Table 1, for the Qiu Er Gou section where calcrete and nodules have been collected in nearby horizons, there is no significant isotopic difference between the two facies. These two components are thus not distinguished in Figure 5a.

The analytical data are rather clustered (our data shown in black and grey on Fig. 5), with some points lying outside of the main field. The data for calcrete and nodules are shown in Figure 5a and those for limestones in Fig. 5b. On the Manas section, the lowest $\delta^{18}\text{O}$ value (19.2‰) was measured for the geodic calcite (sample MC 11-3A, see Fig. 5a), suggesting that this calcite developed under physicochemical conditions distinctly different from the ones prevailing during the main calcification event. On the same section, the Palaeogene nodule MC 11-SED-6 (see Fig. 5a) is somewhat different from the others, displaying lower $\delta^{18}\text{O}$ (20.4‰) and $\delta^{13}\text{C}$ (-6.1‰) values. On the Yaha section, the two Jurassic nodules (YA 12-SED-39 and 41, see Fig. 5a) display extreme $\delta^{13}\text{C}$ values, respectively of -7.6 and -2.0‰. Similarly, the Lower Cretaceous nodule (YA 12-CHIM-1, see Fig. 5a) is also distinguishable from the Upper Cretaceous – Lower Palaeogene calcrete by higher $\delta^{18}\text{O}$ (24.6‰) and $\delta^{13}\text{C}$ (-2.98‰) values. On the Yaha section, while the calcrete displays homogeneous isotopic

compositions, the nodules show wider isotopic ranges and only the Palaeogene nodule YA 12-SED-19A (see Fig. 5a) shows a composition identical to that of the Yaha calcrete.

However, in spite of these differences, the Upper Cretaceous - Lower Palaeogene calcrete and nodules of the Tian Shan display rather homogeneous oxygen and carbon isotope compositions, irrespective of the sampling site. Fluctuations are small in comparison to the range of possible values.

Indeed, most of the Upper Cretaceous - Lower Palaeogene calcrete and nodule $\delta^{18}\text{O}$ values are between 21 and 25‰ (SMOW), while most of the $\delta^{13}\text{C}$ values are between -4 and -6‰ (PDB) (Fig. 5a). On the other hand, the Tian Shan calcrete and nodules differ overall from other calcretes in China (Early Cretaceous in age; Li et al., 2013) by lower $\delta^{13}\text{C}$ values and more homogeneous compositions.

Finally, the Middle Jurassic limestone beds from the Yaha section have highly variable oxygen and carbon isotope composition: $\delta^{18}\text{O}$ values are between 18 and 23‰ (SMOW) and $\delta^{13}\text{C}$ values are between -3.5 and +3.1‰ (PDB), well outside the field of unaltered marine limestones (Fig. 5b). When compared to the lacustrine limestones analysed by Li et al. (2013), the values obtained in the Yaha limestones fall within the field of the SE China and NW China data. We thus interpret those limestones as lacustrine deposits.

6. Discussion

The stable isotope dataset presented here allow us discussing several characteristics of the palaeo-environmental evolution of the Chinese Tian Shan. Indeed, the isotope compositions of the pedogenic carbonates formed in equilibrium with soil or ground water are believed to record a significant part of the original signature of the water, which is itself influenced by several geographical and morphological parameters: continentality, latitude and altitude. Consequently, besides climatic evolution, any topographic changes across space or through time will induce an evolution in the isotopic record.

The Yaha section provides a record of calcification conditions in both the Late Jurassic and the Late Cretaceous - Early Palaeogene times, which helps discussing the evolution of the southern foothills during the Late Mesozoic. The pre-requisite of such analyses is that calcification developed soon after sediment deposition. Although there is no indication for this in the Tian Shan, the nodules were often found in close association with the calcrete (Fig. 3c) and were thus very likely developed

close to the surface. Only the small geodic calcite with an oxygen isotope composition distinct from the whole nodule and calcrete population (sample MC 11-3A) obviously developed in association with some diagenetic event, after the main calcification phase. This shows that the diagenetic circulations that occurred in the series induced isotopic deviation strong enough to be recorded in our data. To a first order, excluding these data (see sample MC 11-3A), it is thus reasonable to discuss the Late Cretaceous - Early Palaeogene palaeo-climatic and palaeo-topographic characteristics, using calcretes and nodules that developed at this stratigraphic level. Finally, some aspects linked to the process of calcification itself can be discussed.

6.1 Isotopic variability within a calcification site

As previously mentioned, numerous processes can influence the isotopic values of pedogenic carbonates (e.g. Ufnar et al., 2008; Charreau et al., 2012; Li et al., 2013), independently of the initial composition of the meteoric water. Some effects are related to the hydrological conditions (evaporation, respiration, seasonality) and influence the oxygen isotopic record. Others are linked to the nature of the palaeo-flora and the relative amount of organic carbon over inorganic carbon hosted in limestones which are mostly of marine origin. It is far beyond the scope of this study to discuss the impact of each of these processes on the isotopic composition of the nodules and calcrete that we measured (the reader is invited to refer to the exhaustive discussion developed in Ufnar et al., 2008; Charreau et al., 2012; Li et al., 2013). Instead, we sampled two neighbouring sites (the Manas and Qiu Er Gou sections) in the northern foothills in order to specify the range of isotopic heterogeneity associated with the calcification processes in our study area. The range of values of the Qiu samples are 21.4 to 24.6‰ ($\delta^{18}\text{O}$) and -4.29 to -6.14‰ ($\delta^{13}\text{C}$), while the range of the Manas samples are 19.2 to 25.3‰ ($\delta^{18}\text{O}$) and -3.75 to -6.08‰ ($\delta^{13}\text{C}$). We consider these ranges as quite comparable.

Each site displays a similar range in $\delta^{18}\text{O}$ values for the Upper Cretaceous - Lower Palaeogene calcrete and nodules ($\delta^{18}\text{O}$ between 22 and 25.3‰ for the Manas site, between 21.4 and 24.6‰ for the Qiu Er Gou site; Table 1). The $\delta^{13}\text{C}$ values are also comparable ($\delta^{13}\text{C}$ between -3.7 and -4.9‰ for the Manas site, between -4.3 and -6.0‰ for the Qiu Er Gou site). The carbonation processes acting at both sites were thus nearly identical, inducing comparable isotopic heterogeneity.

Also, some neo-formed carbonates display specific isotopic compositions in relation to original textural characteristics. The most obvious one is the geodic calcite that developed in the calcrete of the Manas section (MC 11-3A). This calcite has a $\delta^{18}\text{O}$ of 19.2‰ which is lower than that one of its calcrete host (22.2‰). Two main factors can lower the $\delta^{18}\text{O}$ value of carbonate. Firstly, the

$\delta^{18}\text{O}$ value of the fluid may have been 3‰ lower when the geode developed than when the calcrete formed. Secondly, the isotopic fractionation is inversely correlated with temperature. Assuming a constant $\delta^{18}\text{O}$ value of the water from which precipitation occurred, a higher temperature of precipitation during the formation of the geode could result in a lower $\delta^{18}\text{O}$ value of the calcite. We favor this second hypothesis for two reasons. Firstly, lowering the $\delta^{18}\text{O}$ value of the original meteoric water would imply a drastic change in palaeo-climate or palaeo-environmental conditions between the calcrete stage and the geodic stage, a set of conditions that should be recorded by other proxies like the sedimentological record itself and in neighbor areas of Central Asia, which is not the case (e.g. Li et al., 2004). Secondly, a decrease of 3‰ of the $\delta^{18}\text{O}$ value of calcite corresponds to an increase of the temperature of calcite precipitation of about 12-15°C: if calcrete precipitated at 20°C, which corresponds reasonably to pedogenic carbonation conditions, the geode may have precipitated at about 32-35°C. This temperature increase may correspond to burial of the series of a few hundreds of metres within a normal geothermal gradient, a burial that actually occurred in the area during the Cenozoic. Therefore, the geodic calcite developed under late diagenetic conditions rather than under early pedogenic ones.

6.2 Temporal evolution on the southern foothills (Yaha section)

Along the Yaha section, the calcification features formed merely during the Late Jurassic and the Late Cretaceous - Early Palaeogene. These pedogenic carbonates suggest that both periods were characterized by a similar semi-arid climate (e.g. Hendrix et al., 1992; Li et al., 2004; Sun and Wang, 2005). Moreover, the oxygen isotope composition of the neo-formed carbonates is comparable: from 21.2 to 22.6‰ for the Jurassic, and from 21.6 to 23.4‰ for the Early Palaeogene, which is again consistent with similar climatic conditions. In fact, the Late Cretaceous sample YA-12-CHIM-1 displays a $\delta^{18}\text{O}$ value (24.6‰) slightly above the other Yaha samples. As we have only one sample for this intermediate stratigraphic level, we can hardly consider this single value as representative of a regional palaeo-environmental event. It remains that the homogeneity of the stable isotope compositions of the calcification features studied here may indicate that our samples underwent comparable in-situ processes such as evaporation. This homogeneity also suggests that the composition of the rainwater, and by inference also the fundamental parameters evaporation, continentality, latitude, and altitude were likely comparable in the Yaha region for the Late Jurassic and Early Palaeogene periods. In other words, based on our isotopic data no major palaeo-

geographic changes occurred during this period in the Yaha region (confirms earlier data published by Li et al., 2004), i.e. no significant topographic variations occurred either by relief building or destruction. Furthermore, the transgression of the Paratethys Sea from the West into the Tarim Basin during the Late Cretaceous (e.g. Sobel, 1999) had no influence on the conditions of development of pedogenic carbonates in the Yaha section.

In the Yaha region, the isotopic compositions of the Middle Jurassic limestones are different from those of marine limestones of the same age elsewhere in the world (Veizer et al., 1999) (Fig. 5b). It could be argued that these limestones were originally marine and that subsequent diagenesis affected their primary compositions, as known for many carbonate platforms (e.g. Brigaud et al., 2009). However, the fact that $\delta^{13}\text{C}$ of the limestones is highly variable and that the $\delta^{13}\text{C}$ values are not correlated with the $\delta^{18}\text{O}$ values, does not plead for such a diagenetic modification of marine limestones. Indeed, the $\delta^{18}\text{O}$ values of the limestones ranging between 18 and 23‰, the large $\delta^{13}\text{C}$ variation with both positive and negative values and the lack of correlation between the oxygen and carbon signatures are common features of lacustrine carbonates (e.g. Li et al., 2013 and references therein). These authors provided a very detailed discussion on the use of these isotopic signatures in terms of palaeo-environmental reconstructions. In the present work, it would be hazardous to develop such a discussion because of the rather small number of isotopic data. Our inference that the limestone beds are lacustrine is consistent with the silty to sandy continental nature of the series in which they are inter-bedded (see field description above and in Hendrix et al., 1992; Sobel, 1999; Jiang et al., 2008; Heilbronn, 2014).

6.3 Palaeo-topography during the Late Cretaceous - Early Palaeogene period

We have analysed calcification features in three Upper Cretaceous - Lower Palaeogene series in the north, the south and within the Tian Shan Range. Currently, there are two first order differences between the northern and southern sides of the range (Charreau et al., 2012): 1) the cumulative meteoric precipitation rate is much higher on the northern side; i.e., between 100 and 250 mm/year in the southern Junggar Basin than on the southern side of the Tian Shan; i.e., less than 40 mm/year in the Tarim Basin, and 2) the mean $\delta^{18}\text{O}$ value of meteoric water is lower than -8‰ on the northern side and higher than this value on the southern side. If a significant palaeo-topography

existed during the Late Cretaceous - Early Palaeogene period, such an isotopic gradient in meteoric water compositions likely existed and should have been recorded in pedogenic carbonates.

From the lithological point of view, our field observation showed that calcrete features developed mainly in conglomerates, i.e. on rocks with comparable porosity and permeability properties. Also, a semi-arid climate prevailed over the whole study area in order to develop pedogenic carbonates (e.g. Hendrix et al., 1992). It is then reasonable to think that the prominent variable that could have been recorded by the carbonates developed in the Upper Cretaceous - Lower Palaeogene series, is the initial isotopic composition of the meteoric water, a property that depends primarily on the origin of atmospheric masses (e.g. Charreau et al., 2012). In Table 1 and Figure 5, there is no significant isotopic difference between the Upper Cretaceous - Lower Palaeogene pedogenic carbonates of the northern, central and southern parts of the present-day Tian Shan Range. The interpretation that immediately comes out is that the meteoric waters were identical in the three regions.

The easier way to reach identical values is to consider that these meteoric waters originated from the same oceanic reservoir and underwent the same atmospheric evolution. The Paratethys sea in the West-Tarim Basin (e.g. Dercourt et al., 1993; Mao and Norris, 1988) could be a possible source of these meteoritic waters. Regardless of the marine source of meteoric waters, there is no isotopic evidence for the existence of any orographic barrier between the studied sections at the time of calcification. If the latter rapidly followed deposition, there was no significant relief in that area at the Cretaceous - Palaeogene boundary.

7. Conclusion

Pedogenic carbonates (calcrete and nodules) and limestone beds have been identified and analysed for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ on two main stratigraphic levels of the Tian Shan Mesozoic-Cenozoic series (Late Jurassic and Late Cretaceous - Early Palaeogene). The limestone beds are restricted to the Middle Jurassic series of the southern foothills and are probably of lacustrine origin. The oxygen and carbon isotope compositions of the pedogenic carbonates are homogeneous in both the southern and northern foothills of the Tian Shan Range, as well as in the intramontane Bayanbulak Basin. Palaeo-environmental conditions of pedogenic carbonates development were thus identical on a regional scale during the Late Cretaceous - Early Palaeogene period. No significant palaeo-relief

existed by that time in the area, implying that the actual Tian Shan topography mostly results from the Cenozoic reactivation event.

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Table caption

Table 1- Oxygen and carbon isotope composition of the pedogenic and lacustrine carbonates from the Tian Shan Range.

Figure caption

Fig. 1- a) Schematic tectonic map of Asia and location of the Tian Shan Range in Central Asia. b) Map of the Tian Shan area (DEM, G-TOPO-30, Mercator projection) showing the five sampling sites: the Wusu, Manas and Qiu Er Gou sections on the northern foothills, the Bayanbulak section in the Bayanbulak intramontane basin, and the Yaha section on the southern foothills of the range. Section location and GPS coordinates are given in Table 1.

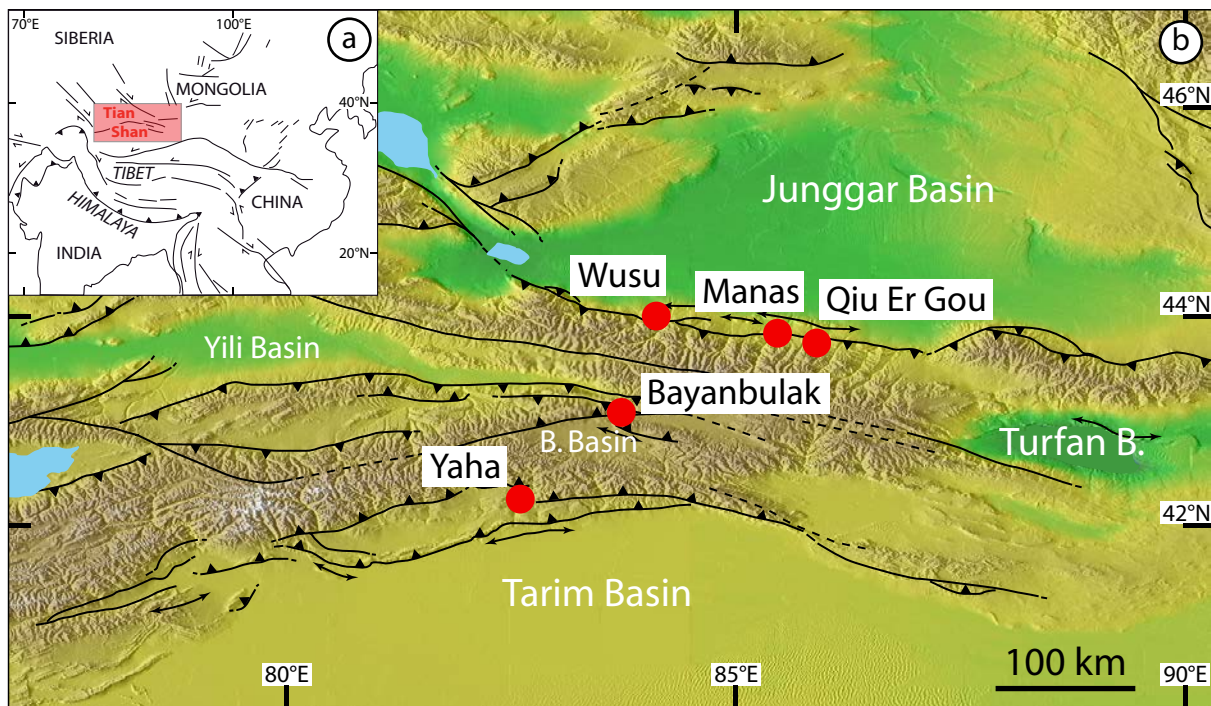
Fig. 2- Synthetic figure showing (top) the present-day topographic profile through the Tian Shan Range, together with the five sampling sites (black dots; W. = Wusu; M. = Manas; Q. = Qiu Er Gou; B. = Bayanbulak; Y. = Yaha). (left) General lithological column (C = Clay, s = Silt, Sf = Fine sand, Sm = Medium sand, Sc = Coarse sand, g = Gravel, P = Pebble, B= Boulder) derived from our field observations of the Middle Jurassic to Palaeocene series from the Yaha section (southern foothills). This column shows the general trends and types of Mesozoic deposits around the Tian Shan Range and is consistent with the work of Hendrix et al. (1992). The correspondence between lithological limits and stratigraphic ages relies on the available geological maps, previous studies (Hendrix et al., 1992) and our field work (J1 = Early Jurassic; J2 = Middle Jurassic; J3 = Late Jurassic; K1 = Early Cretaceous; K2 = Late Cretaceous; P = Palaeocene). (right) Type and stratigraphic position of the neo-formed carbonates.

Fig. 3- Outcrop views. a) Landscape view of the sampled 8 metre-thick massive calcrete, which runs over tens of kilometres laterally (Late Cretaceous – Early Palaeocene, top of the Manas Section). b) Detailed view of the massive conglomeratic calcrete (hammer for scale). The boundaries of the carbonate body correspond to lithological limits between sandy layers and gravel-rich layers. c) Transition from nodule-rich layers into a massive calcrete (Late Cretaceous – Early Palaeocene, Yaha

Section; hammer for scale). d) Decimetric grey to white limestone layers. These layers are not easily distinguishable solely from their colour.

Fig. 4- Examples of studied samples. a) Conglomeratic massive calcrete (sample MC 11 4; Late Cretaceous – Early Palaeocene; Manas section). b) Centrimetre-sized calcareous nodules (sample B 11 SED 1; Early Palaeocene; Bayanbulak section). c) Limestone, locally with shales or terrigenous input (sample YA 12 SED 1; Middle Jurassic; Yaha section).

Fig. 5- $\delta^{13}\text{C}$ (PDB) vs. $\delta^{18}\text{O}$ (SMOW) diagram of the pedogenic carbonates (a) and the limestones (b) of the Tian Shan Range. The isotopic compositions of Lower Cretaceous pedogenic carbonates and limestones from China are shown for comparison (data from Li et al., 2013). The field of Jurassic marine limestones of the world is indicated in grey (Veizer et al., 1999).



Heilbronn et al., 2014 - Fig. 1

Heilbronn et al., 2014 - Fig. 2

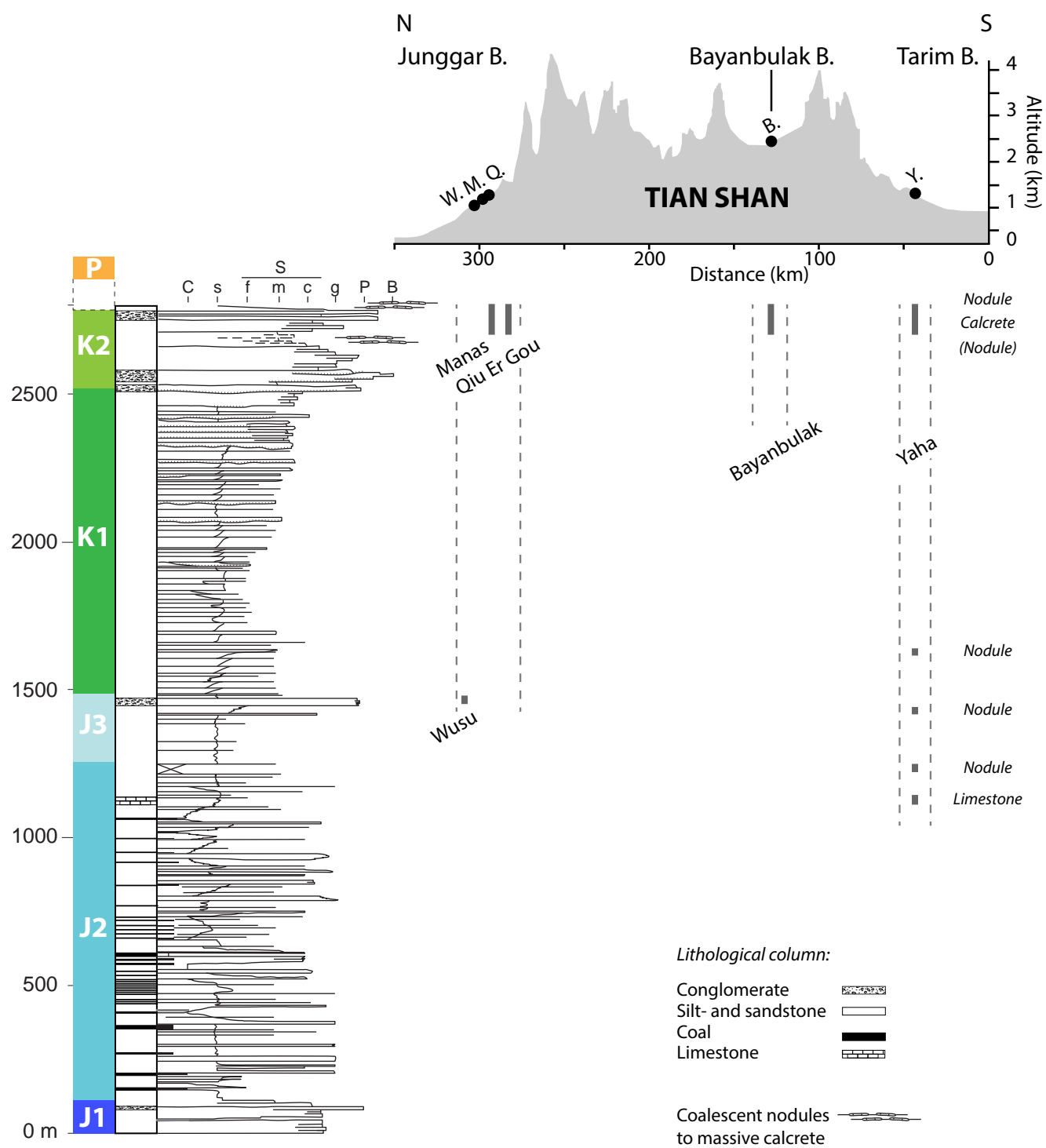


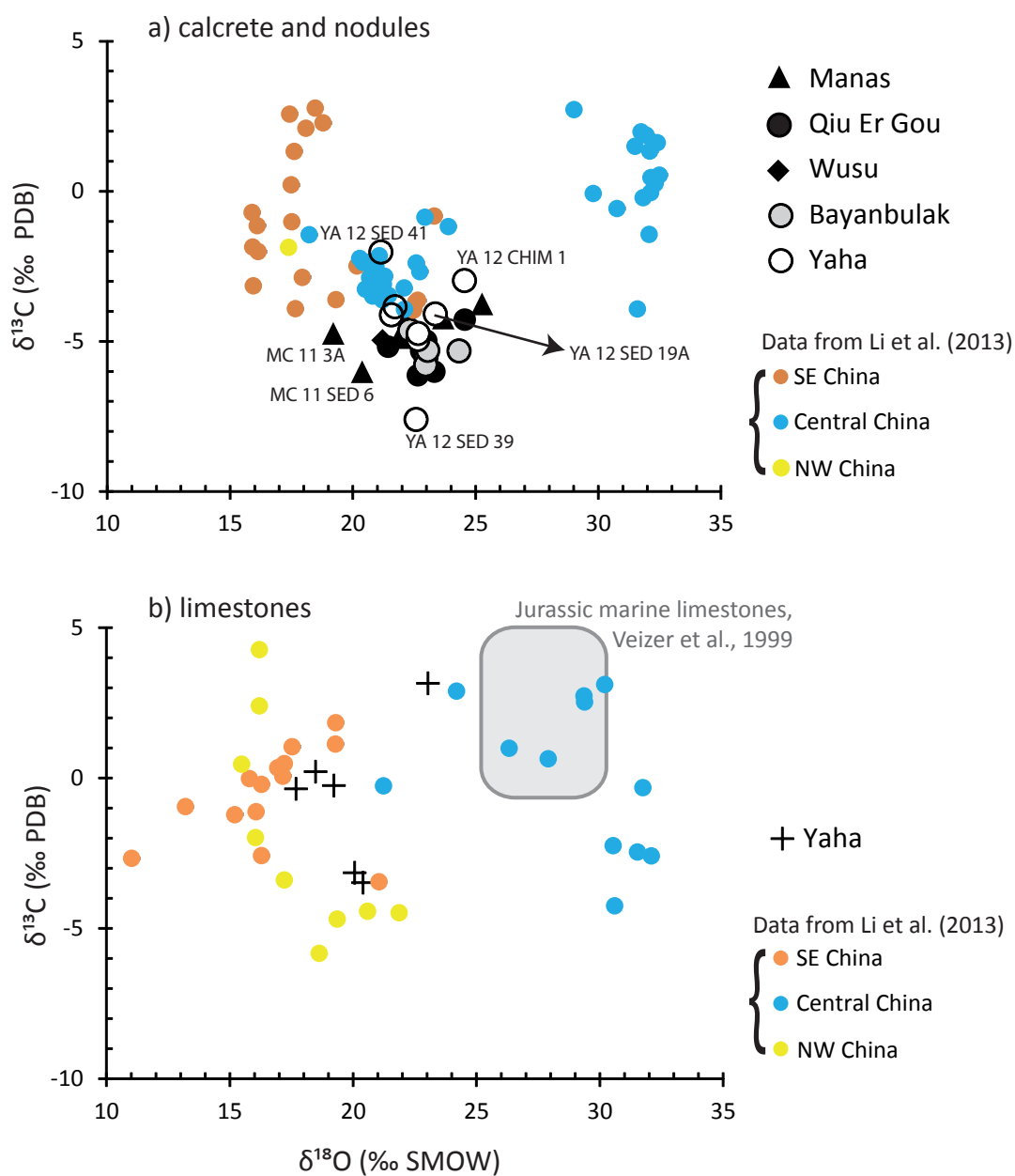


Fig. 3 - Outcrop-views.





Fig. 4 - Examples of studied samples.



Heilbronn et al., Figure 5

Table 1

			Coordinates		Calcite	
Sample	Type	Formation	Latitude	Longitude	δ ¹⁸ O	δ ¹³ C
WUSU						
WE 12 SED 1	Calcar. nodules	Jurassic (J3)	44° 8'20,90"N	84°30'19,70"E	21.2	-4.96
MANAS						
MC 11 SED 6	Calcar. nodules	Palaeogene	43°57'55,08"N	85°48'28,08"E	20.4	-6.08
MC 11 5	Calcrete	Cret./Palaeogene	idem		25.3	-3.75
MC 11 4	Calcrete	Cret./Palaeogene	idem		23.7	-4.22
MC 11 3C	Calcrete (matrix)	Cret./Palaeogene	idem		22.2	-4.05
MC 11 3B	Calcrete (vein)	Cret./Palaeogene	idem		22.1	-4.87
MC 11 3A	Calcrete (geode)	Cret./Palaeogene	idem		19.2	-4.73
MC 11 2	Calcrete	Cret./Palaeogene	43°57'42,72"N	85°48'27,60"E	22.0	-4.67
QIU ER GOU						
HU 12-SED 12	Calcar. nodules	Cret./Palaeogene	43°52'3,00"N	86°23'32,30"E	21.4	-5.19
HU 12-SED 11	Calcar. nodules	Cret./Palaeogene			23.3	-6.02
HU 12-SED 10	Calcrete	Cret./Palaeogene			24.6	-4.29
HU 12-SED 9	Calcrete	Cret./Palaeogene			23.0	-4.99
HU 12-SED 8	Calcar. nodules	Cret./Palaeogene			22.8	-5.33
HU 12-SED 7	Calcrete	Cret./Palaeogene	43°52'2,40"N	86°23'29,50"E	22.7	-6.14
BAYANBULAK						
B 11 SED 1	Calcar. nodules	Palaeogene?	43° 4'23,52"N	84°17'6,66"E	22.3	-4.65
B 11 SED 06	Calcrete	Cret./Palaeogene?	idem		23.1	-5.31
B 11 SED 02	Calcrete	Cret./Palaeogene?	idem		23.0	-5.79
B 11 SED 05	Calcrete	Cret./Palaeogene?	idem		24.3	-5.32
B 11 SED 04	Calcrete	Cret./Palaeogene?	43° 4'21,96"N	84°17'5,52"E	23.0	-5.53
YAHA						
YA 12 SED 19 A	Calcar. nodules	Palaeogene	42° 4'52,32"N	83°15'48,20"E	23.4	-4.10
YA 12 SED 18	Calcrete	Cret./Palaeogene	42° 4'56,30"N	83°15'48,20"E	22.7	-4.75
YA 12 SED 17	Calcrete	Cret./Palaeogene	42° 4'57,15"N	83°15'48,61"E	22.7	-4.95
YA 12 SED 13	Calcrete	Cret./Palaeogene	42° 4'56,84"N	83°15'50,13"E	21.6	-4.13
YA 12 SED 16	Calcrete	Cret./Palaeogene	42° 4'57,32"N	83°15'50,13"E	21.7	-3.86
YA 12 CHIM 1	Calcar. nodules	Cretaceous (K1)	42° 5'47,80"N	83°16'40,10"E	24.6	-2.98
YA 12 SED 41	Calcar. nodules	Jurassic (J3)	42° 6'52,30"N	83°17'18,50"E	21.2	-2.03
YA 12 SED 39	Calcar. nodules	Jurassic (J2-J3)	42° 6'57,90"N	83°17'18,10"E	22.6	-7.61
YA 12 SED 38	Limestone	Jurassic (J2)	idem		20.1	-3.16
YA 12 SED 37	Limestone	Jurassic (J2)	idem		19.2	-0.26
YA 12 SED 36	Limestone	Jurassic (J2)	idem		18.5	0.20
YA 12 SED 34	Limestone	Jurassic (J2)	idem		23.0	3.14
YA 12 SED 33	Limestone	Jurassic (J2)	42° 7'3,37"N	83°17'7,09"E	17.7	-0.37
YA 12 SED 1	Limestone	Jurassic (J2)	42° 7'32,40"N	83°14'10,40"E	20.4	-3.49

CHAPITRE 5 – DISCUSSION GENERALE, CONCLUSIONS ET PERSPECTIVES :

Evolution mésozoïque de la chaîne du Tian Shan chinois dans le contexte géodynamique de l'Asie Centrale

L'objectif de cette thèse a été de reconstruire l'évolution de la chaîne du Tian Shan chinois durant le Mésozoïque, ainsi que de restituer la topographie de premier ordre dans la région. L'originalité de ce travail réside dans la combinaison d'une étude sédimentologique avec des résultats de thermochronologie basse-température. Ainsi, le paradoxe du Jurassique supérieur opposant les données de sédimentologie à celles de thermochronologie basse-température (cf. Introduction, Chapitre 2 – Sédimentologie) a pu être résolu. Ceci souligne l'absolue nécessité de d'associer plusieurs méthodes pour étudier l'enregistrement de l'évolution d'une chaîne de montagnes.

Au Trias supérieur, la chaîne du Tian Shan enregistre toujours une phase d'exhumation et les reliefs sont encore significatifs (e.g. Dumitru et al., 2001 ; Jolivet et al., 2013), tout comme en Asie du Sud-Est et au Tibet (e.g. Sone et Metcalfe, 2008 ; Roger et al., 2010). Les reliefs du Tian Shan résultent de la phase compressive engendrée par l'accrétion du bloc de Qiangtang au sud, au Trias inférieur, qui a donné lieu à l'orogénèse Cimmérienne (e.g. Roger et al., 2008, 2010) (Fig. 5.1 et 5.2.a). Plus loin au nord-est de l'Asie, la zone de subduction du Mongol-Okhotsk est active, probablement depuis le Dévonien. L'Océan se ferme en ciseaux et de manière progressive, entraînant une extension d'arrière-arc au sein du Bloc Sibérie (e.g. Barbarin, 1999 ; Donskaya et al., 2013). L'Asie centrale ne semble pas être affectée par cette extension, et la région du Mongol-Okhotsk est marquée par la mise en place de batholithes (Khentey par exemple) indiquant la position de la marge active (Donskaya et al., 2013 et références associées).

Cette thèse [Tian Shan]			Synthèse bibliographique [Asie Centrale]			Asie en général	
	Topo.	Tecto.		Topo.	Tecto.		
	Relief/exhumation	Réactivation (compression)	Goby Altay, Sayan, Nord-Tibet Pamir	Exhumation	Compression		Ouv
Crête	Très faible relief	Transpression (mouvements décrochants dextres)				Collision du Kohistan, l'arc du Drias et Bloc Afghan	
ation raire Tarim	Relief/réactivation localisée			Stabilité thermique (Altay) + pénplaine		Paroxysme de l'extension dans la région du bassin S-Caspien	Inv
	Exhumation du Tian Shan ouest		MCC et horst/graben en Mongolie Bassins extensifs en Chine NE		Extension NO-SE		
	Très faible relief/Exhumation lente & continue	Inversion des bassins				Collision de Lhasa	Ferm
				J/K = refroidissement et exhumation	J/K = réactivation Altay	J/K = changement cinématique plaque Izanagi => extension; J/K = orogénèse Yanshan?	
Cônes luviaux	Faible relief / Exhumation très faible	Extension	Compression: Ouest-Ts, N-Altay, Sayan (sauf E-Sayan) Extension: Patom, Transbaikai		Activité tectonique modérée	Ouverture proto-bassin S-Caspien	
n du r	Pénplaine / Faible exhumation	Transension	Du Nord-Tibet au sud-est de la Sibérie, + Patom	Pénplaine, dépôt de charbon			
canisme spots de charbon	Relief Exhumation	Transension (extension dans Turfan & Tarim)	Mongolie-ouest, Goby Altay, Sayan, Nord-Tibet	Erosion reliefs du Paléozo. sup/exhumation			
	Relief Exhumation	Compression			Compression	Fin de la collision de Qiangtang	Sub M-Okt

Fig. 5.1 : Tableau de synthèse des évènements géodynamiques affectant l'Asie Centrale durant le Mésozoïque. Les évènements surlignés en vert indiquent les évènements montrant un régime tectonique concordant avec les observations dans le Tian Shan, et ceux surlignés en orange un régime contradictoire.

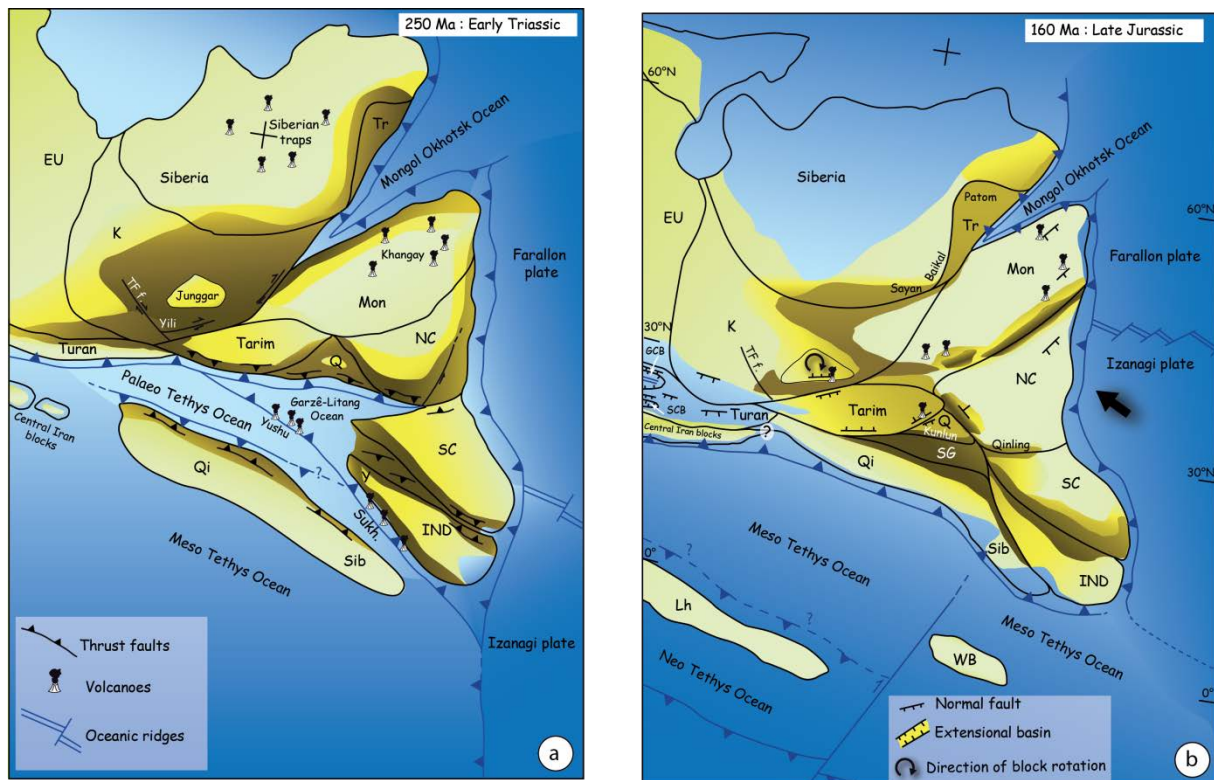


Fig. 5.2 (a, b) : Cartes paléogéographiques de l'Asie, indiquant le relief supposé. a = Trias inférieur (250 Ma), b = Jurassique supérieur (160 Ma) (Jolivet, in review)

EU = plaque européenne ; K = Kazakhstan ; Tr = Transbaikal ; Mon = Mongolie ; NC = Chine du Nord ; SC = Chine du Sud ; Q = Qaidam ; SG = Songpan-Garzê ; Qi = Qiangtang ; Sib = Sibumasu ; IND = Indochine ; Lh = Lhassa ; WB = Burma ouest ; Kh = Kohistan.

Dans le Tian Shan, au Jurassique inférieur la topographie tardi-paléozoïque est progressivement érodée et le socle continue d'être exhumée (Dumitru et al., 2001 ; Jolivet et al., 2010). Cette période semble dominée par un régime transtensif accompagné d'une activité décrochante le long des principales failles (faille du Nord Tian Shan par exemple). Les Bassins de Turfan et du Tarim quant à eux, sont en extension (e.g. Allen et al., 1991a ; Shao et al., 2003). Ailleurs en Asie Centrale, cette phase d'exhumation est également enregistrée (Nord-Tibet : Jolivet et al., 2001 ; Mongolie ouest : Jolivet et al., 2007 ; Gobi Altay : Vassallo et al., 2007 ; Sayan : De Grave et Van den haute, 2002 ; Jolivet et al., 2013b) et l'érosion des reliefs paléozoïques supérieurs se poursuit. Le Jurassique inférieur pourrait donc correspondre à une période de transition, qui suit l'orogénèse Cimmérienne.

Au Jurassique moyen, les taux d'érosion observés en Asie Centrale de manière générale sont très faibles et s'accompagnent de la formation d'une pénéplaine, non seulement dans le Tian Shan, mais également dans le reste de l'Asie Centrale (du Nord-Tibet jusqu'au sud-est de la Sibérie) (e.g. Dumitru et al., 2001 ; Jolivet et al., 2001 ; Vassallo et al., 2007 ; Jolivet et al., 2007, 2010 ; Glorie et al., 2010). Depuis le début du Jurassique, le bloc du Junggar est affecté par une rotation de sens horaire, créant des paléo-hauts topographiques au sein du bassin (Hou et al., 2009) et permettant l'expression simultanée de marqueurs de l'extension et de la compression sur les différents bords du bassin (e.g. Allen & Vincent, 1997 ; Liu et al., 2006). Par conséquent, nous supposons un régime tectonique en transtension dans le Tian Shan, et s'il existe de l'activité tectonique dans le reste de l'Asie Centrale, elle n'entraîne pas d'exhumation mesurable. A l'ouest, l'extension d'arrière-arc de la subduction de la Méso-Téthys entraîne l'ouverture du proto-Bassin Sud-Caspien (e.g. Brunet et al., 2003), indiquant un régime extensif qui n'est pas contradictoire avec le régime tectonique observé dans la région du Tian Shan. Cependant, au nord-est de l'Asie, cette période correspond à la fermeture de l'Océan du Mongol-Okhotsk et par conséquent au début de la « collision » entre la Sibérie et la Mongolie (e.g. Zorin, 1999 ; Kravchinsky et al., 2002a,c ; Cogné et al., 2005 ; Jolivet et al., 2009, 2013a ; Donskaya et al., 2013). Cet événement ne semble pas affecter la région du Tian Shan, ni le reste de l'Asie Centrale à cette époque. En revanche, la phase terminale de la fermeture de l'Océan (au Jurassique supérieur) semble entraîner le début de l'exhumation de la chaîne de l'Altay, dont la phase majeure de réactivation est enregistrée à la limite Jurassique-Crétacé (e.g. De Grave et al., 2007, 2008). Cette réactivation est induite par le régime tectonique compressif lié à la collision entre les continents Sibérie et Amur (Chine du Nord + Mongolie) dans la région du Mongol-Okhotsk. L'enregistrement des événements liés à la « collision » du Mongol-Okhotsk reste encore incomplet et largement débattu. L'absence d'une phase d'exhumation dans le sud-est de la chaîne des Sayan est (Jolivet et al., 2013b), ainsi que l'absence de discordance et de variation majeure des environnements de dépôts dans le Bassin Ouest-Sibérien et dans la région Transbaïkale (e.g. Davies et al., 2010 ; Jolivet et al., 2009, 2013a) ne permettent pas de conclure à une large orogénèse.

Au Jurassique supérieur, des reliefs n'engendrant pas d'exhumation décelable par les analyses traces de fission sur apatite, se forment dans la région du Tian Shan (cf. Chapitre 2). Un équivalent des épais dépôts conglomératiques indiquant l'érosion de ces reliefs (décrit dans le Tian Shan chinois, Formation de Kalazha cf. Fig. 2.3, 2.4 et 2.9 et Fig. 2.10, 2.11 et 2.16) a été observé plus loin à l'ouest, dans le Tian Shan kirghize, au nord du Bassin de Ferghana (Fig. 5.3 et 5.4, Bourquin et al., in review). Sachant que cette période est caractérisée par une très faible exhumation dans l'ensemble de la chaîne (e.g. Dumitru et al., 2001 ; Jolivet et al., 2010 ; Glorie et al., 2010), nous

interprétons le régime tectonique de cette époque comme globalement extensif (cf. Chapitre 2). Il s'accompagne d'une aridification à l'échelle de la chaîne, qui s'est mise en place progressivement (e.g. Allen et al., 1991b; Hendrix et al., 1992; Parrish et al., 1993; Shao et al., 2003; Li et al., 2004; Eberth et al., 2001; Pan et al., 2013). A l'ouest, l'ouverture du proto-bassin Sud-Caspien se poursuit, de même que la fermeture de l'Océan Mongol-Okhotsk à l'est (Figure 5.2.b). Au Tian Shan, aucune réactivation comparable à celle de la chaîne de l'Altay n'est enregistrée et les surfaces de pénéplaine formées au Jurassique moyen, conservées aujourd'hui encore attestent d'une faible exhumation de la chaîne depuis. Dans la région de l'Altay, la réactivation tardi-Jurassique est suivie d'une phase de calme tectonique allant de la fin du Crétacé jusqu'au début du Néogène, et est accompagnée de la formation d'une surface de pénéplaine (e.g. De Grave et al., 2008). Ceci démontre bien l'hétérogénéité des déformations affectant l'Asie Centrale durant le Mésozoïque.

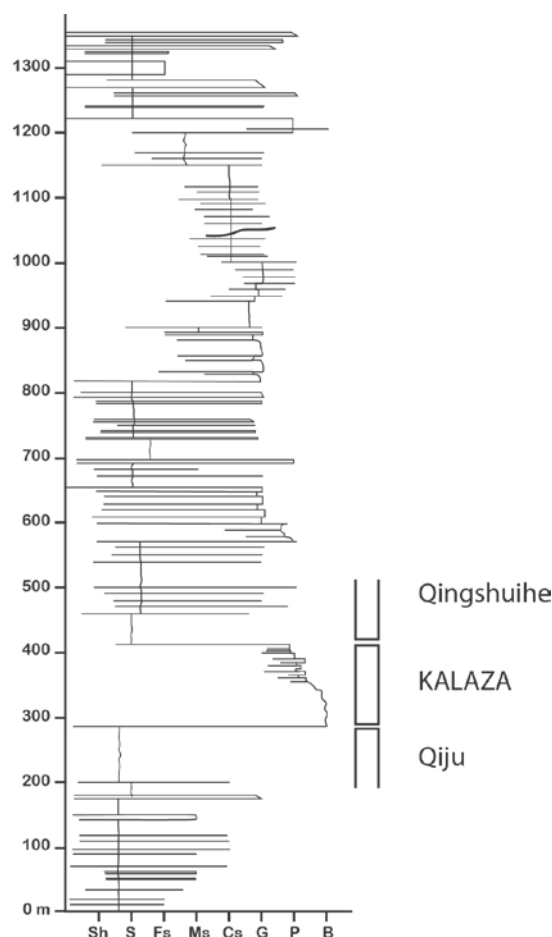


Figure 5.3 Log représentant les séries mésozoïques au nord-est du Bassin de Ferghana (Kirghizstan) et indiquant la présence d'épais dépôts de cônes alluviaux à la fin du Jurassique – début du Crétacé, équivalents de la Formation de la Kalazha, observée à l'est dans le Tian Shan chinois (cf. Fig. 2.3, 2.4 et 2.9 et Fig. 2.10, 2.11 et 2.16) (Bourquin et al., accepted).





Figure 5.4 (a, b et c) Photos des dépôts de cônes alluviaux équivalents à la Formation de la Kalazha, observée au nord-est du Bassin de Ferghana (Kirghizstan) (Bourquin et al., in review). a. Vue générale de la base de la formation. – b. et c. Dépôts conglomératiques, claste-supportés (b.) ou à mégarides (c.).

Dans le nord-est de la Chine, Faure et al. (2012) observent des évidences de la collision des blocs Chine du Nord et Alashan, sur la marge nord du Bloc Chine du Nord et la formation de la chaîne Yanshan- Yinshan à la fin du Jurassique et au début du Crétacé. Cette orogénèse ne semble pas affecter la région du Tian Shan.

A la limite entre le Jurassique et le Crétacé, un changement de cinématique de la Plaque Izanagi (d'orthogonale à fortement oblique, ainsi qu'une accélération de la vitesse de convergence, e.g. Maruyama et Send, 1986; Maruyama et al., 1997) entraîne le retrait du panneau plongeant de cette plaque et donc l'extension arrière-arc dans la plaque continentale eurasiatique (Watson et al., 1987 ; Traynor et Sladen, 1995 ; Ratschbacher et al., 2000 ; Ren et al., 2002 ; Yang et al., 2007) (Fig. 5.5.a). Pour avoir de telles conséquences et être le moteur de l'extension dans une large zone de l'Asie du nord-est, ce mécanisme peut être accompagné d'autres mécanismes lithosphériques (e.g. érosion thermique de la lithosphère, amincissement de la lithosphère) (e.g. Ren et al., 2002 ; Meng et al., 2003). Cependant, l'hypothèse la plus plausible expliquant ce régime tectonique, serait plutôt liée au régime thermique de la croûte supérieure. En effet, celle-ci est anormalement chaude suite à l'intense activité magmatique qu'elle a subie précédemment et durant le Crétacé. Ceci a donc permis la mise en place d'un régime extensif très distribué à l'échelle de l'Asie orientale (Daoudene, 2011). Ainsi, dès le Jurassique supérieur, la formation de bassins extensifs a eu lieu dans le nord-est de la Chine (e.g. Charles, 2010), ainsi que la formation de horsts, grabens et de dômes métamorphiques au Crétacé inférieur en Mongolie (e.g. Daoudene, 2011).

Au Crétacé inférieur, une exhumation lente et continue est enregistrée par la thermochronologie basse-température dans le Tian Shan et va de pair avec une topographie relativement plane (Fig. 5.1). Cette période correspond également au début de la réactivation des bassins (Junggar, par exemple) (e.g. Yang et al., 2013), qui est généralement associée au régime convergent affectant l'ensemble de la marge sud-asiatique, marquée par le début de l'accrétion du bloc de Lhasa (e.g. Kapp et al., 2005, 2007). Pourtant, l'enregistrement de la collision de Lhasa est hétérogène en Asie. L'est du Tibet et la marge sud du bassin du Tarim, des régions situées à proximité de la zone de collision, n'enregistrent que très peu de déformation en réponse à cette collision (e.g. Jolivet et al., 2001 ; Roger et al., 2008, 2010, 2011). Dans le Tian Shan chinois, aucune phase d'exhumation n'a été identifiée au Crétacé inférieur (e.g. Dumitru et al., 2001 ; Jolivet et al., 2010). En revanche, une phase d'exhumation continue est enregistrée dans le Tian Shan ouest au cours du Crétacé et semble être en réponse de la collision du Bloc de Lhasa (De Grave et al., 2007). Cette dichotomie dans l'enregistrement de la déformation observée entre les deux zones de la chaîne pourrait être liée à des mouvements de rotation horaire? du Bloc du Tarim, similaires à ceux observés durant le Tertiaire. Dans le Tian Shan chinois, l'absence d'exhumation est associée à des séries sédimentaires déposées dans des milieux calmes au Crétacé inférieur. Une limite abrupte les sépare des conglomérats du Jurassique supérieur (Formation de Kalazha) et indique un changement rapide et relativement brutal. Le Crétacé inférieur correspond donc à une période charnière faisant suite à l'extension généralisée du Jurassique supérieur et précédant la réactivation au Crétacé supérieur. Par conséquent, il est possible d'imaginer que la collision de Lhasa ait mis fin à la phase d'extension généralisée, enregistrée dans le Tian Shan chinois.

La fermeture de l'Océan Mongol-Okhotsk reste également l'objet de nombreuses discordes, dues à l'hétérogénéité de son enregistrement. Malgré des signes apparents d'une phase d'exhumation au Jurassique supérieur - Crétacé inférieur (thermochronologie et de sédimentologie) enregistrée dans les régions des Sayan, du Transbaikal et de la chaîne de Patom (e.g. Le Heron et al., 2008 ; Van der Beek et al., 1996 ; Jolivet et al., 2009 ; 2013b), ainsi que dans le nord-est de l'Altay (De Grave and Van den haute, 2002 ; De Grave et al., 2008), aucune observation géologique (tectonique, métamorphisme ou sédimentologie par exemple) ne montre d'argument fiable pour une collision entre la Sibérie et la Mongolie (cf. Chapitre 1) (e.g. Zorin, 1999 ; Jolivet et al., 2009 ; Davies et al., 2010 ; Jolivet et al., 2013b). Par conséquent, il est incertain que la fermeture de l'Océan Mongol-Okhotsk ait réellement affecté le régime tectonique en Asie Centrale à cette époque. En revanche, le chemin de migration des pôles de référence des blocs Sibérie et Amur, impliqués dans cette collision, ainsi que la reconstitution des rotations de ces blocs à partir de données de paléomagnétisme issues

des deux blocs, impliquent réellement des processus de collision (e.g. Cogné et al., 2005). Ces données impliquent une marge d'erreur de l'ordre de plusieurs centaines de kilomètres et ne permettent pas encore de disposer d'un enregistrement précis des événements liés à la fermeture de l'Océan du Mongol-Okhotsk.

Au Crétacé supérieur, l'exhumation lente et continue se poursuit dans l'ensemble du Tian Shan, accompagnée d'une topographie positive et enregistrée par les analyses traces de fission sur apatite (e.g. Dumitru et al., 2001 ; Jolivet et al., 2010 ; De Grave et al., 2013). Dans le reste de l'Asie Centrale, le régime thermique est stable et montre un refroidissement lent par érosion (par exemple dans l'Altay) (e.g. De Grave et al., 2008 ; Glorie et al., 2012a). A l'ouest, l'ouverture du Bassin Sud-Caspien atteint son paroxysme. L'extension dans cette région, n'est plus corrélée à la dynamique de l'Asie Centrale depuis la collision du Bloc de Lhassa. Plus loin à l'est, une phase d'inversion tectonique est observée en Mongolie (Daoudene, 2011). La plaque Izanagi est entièrement subduite dès le début du Crétacé supérieur (e.g. Maruyama & Send, 1986 ; Maruyama et al., 1997) et un magmatisme significatif affecte l'est de la Chine (e.g. Charles, 2010) (Fig. 5.5.a). Par conséquent, l'extension tardi-Mésozoïque a considérablement diminué au Crétacé supérieur, avant de reprendre au Tertiaire.

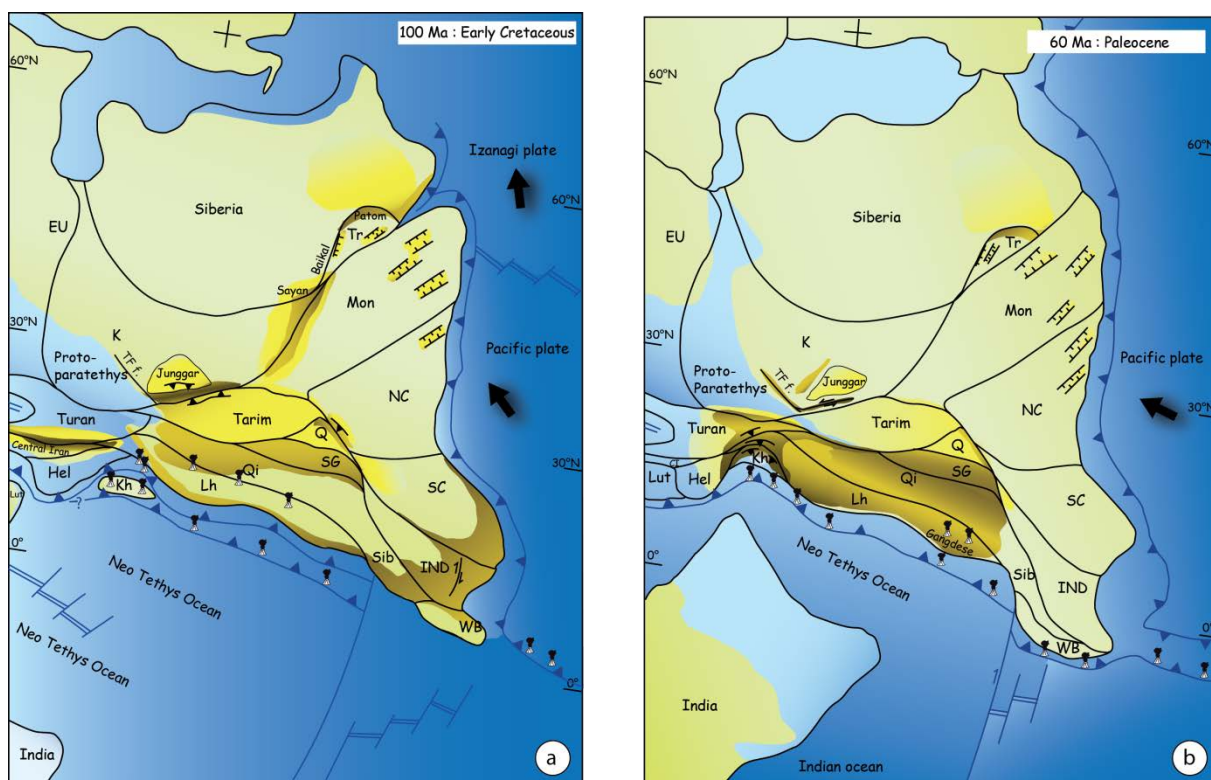


Fig. 5.5 (a, b) : Cartes paléogéographiques de l'Asie, indiquant le relief supposé. a = Fini-Crétacé inférieur (100 Ma), b = Paleocene (62 Ma) (Jolivet, in review)

EU = plaque européenne ; K = Kazakhstan ; Tr = Transbaikal ; Mon = Mongolie ; NC = Chine du Nord ; SC = Chine du Sud ; Q = Qaidam ; SG = Songpan-Garzê ; Qi = Qiangtang ; Sib = Sibumasu ; IND = Indochine ; Lh = Lhassa ; WB = Burma ouest ; Kh = Kohistan.

A la limite Crétacé – Tertiaire, les reliefs présents le long des principales failles dans l'ensemble du Tian Shan, ainsi que l'exhumation enregistrée par la thermochronologie basse-température, indiquent une réactivation localisée le long de ces structures (e.g. Wang et al., 2008 ; Wang et al., 2009 ; Jolivet et al., 2010 ; De Grave et al., 2013 ; Macaulay, 2013), ainsi qu'un régime en transpression, avec des mouvements décrochants dextres. Cette réactivation est généralement associée à l'accrétion de blocs tel que le Kohistan ou l'arc du Dras sur la marge sud de l'Asie (e.g. Hendrix et al., 1992; Jolivet et al., 2010). De plus, la déformation peut s'étendre aisément vers l'est, grâce à la présence du bord libre constitué par la marge pacifique, qui favorise l'extension de la lithosphère dans la zone Sibérie-Mongolie-Chine du Nord (Jolivet et al., 2010) (Fig. 5.5.b).

La topographie due à la réactivation crétacée est probablement érodée au début du Paléocène, permettant la formation d'un niveau de calcrète de plusieurs mètres d'épaisseur dans la zone du Tian Shan (Chapitre 4 – Géochimie). Grâce aux outils isotopiques, nous avons mis en évidence l'absence de relief significatif depuis la fin du Mésozoïque dans la région du Tian Shan, ce qui est confirmé par la conservation d'une surface plane (e.g. De Grave et al., 2008 ; Jolivet et al., 2010 ; Chapitre 3 – Synthèse AFT & Chapitre 4 - Géochimie) et la faible exhumation associée (cf. Chapitre 3 – Synthèse AFT).

Au sud-ouest du Tian Shan, la zone du Pamir commence à se déformer à l'Eocène, sans influence significative pour la région du Tian Shan chinois à cette période (e.g. Schwab et al., 2004 ; Bershaw et al., 2012).

Ceci est suivi par une large réactivation de la chaîne à partir du Miocène, à l'origine des reliefs significatifs actuels (e.g. Dumitru et al., 2001 ; Bullen et al., 2003 ; Sobel et al., 2006a, 2006b ; Wang et al., 2009 ; De Grave et al., 2012, 2013 ; Macaulay et al., 2013 ; Macaulay, 2013). Cette phase d'exhumation est également observée dans le Goby Altay, dans l'Altay, dans les Sayan et dans le Nord-Tibet (e.g. Jolivet et al., 2001 ; De Grave et al., 2008 ; Jolivet et al., 2009 ; Glorie et al., 2012a ; Jolivet et al., 2013b). Cet évènement est une réponse à la collision entre l'Inde et l'Asie 30 Ma plus tôt (e.g. Avouac et al., 1993a ; Charreau et al., 2005, 2009b).

Perspectives

Ce travail ayant une fois de plus montré la nécessité de combiner différentes approches dans l'étude de l'évolution d'un relief, il serait pertinent de poursuivre cette démarche pour mieux connaître les reliefs mésozoïques d'Asie Centrale, mais surtout mieux comprendre l'enregistrement des principales phases d'activité tectonique. Dans un premier temps, l'étude des dépôts mésozoïques du Tian Shan ouest combinés aux données géochronologiques (voir thèse en cours, E. De Pelsmaeker), permettra sans doute la réalisation d'une synthèse à l'échelle de l'ensemble du Tian Shan, ainsi que d'éventuelles reconstitutions paléotopographiques et paléogéographiques plus précises.

L'association de la sédimentologie et de la thermochronologie dans la région des Kunlun, livrerait sans doute des contraintes sur l'enregistrement de la collision du Bloc de Lhasa au Crétacé inférieur et permettrait de mieux comprendre ses répercussions dans le reste de l'Asie Centrale. De la même manière, une étude similaire sur la région du Mongol-Okhotsk alimenterait le débat sur les modalités de fermeture de l'océan et de la collision. Par ailleurs, l'influence du climat, n'est certes pas dominante, mais joue un rôle certain ne serait-ce que dans l'enregistrement de l'activité tectonique et apporterai des contraintes pertinentes, particulièrement dans les reconstitutions paléogéographiques.

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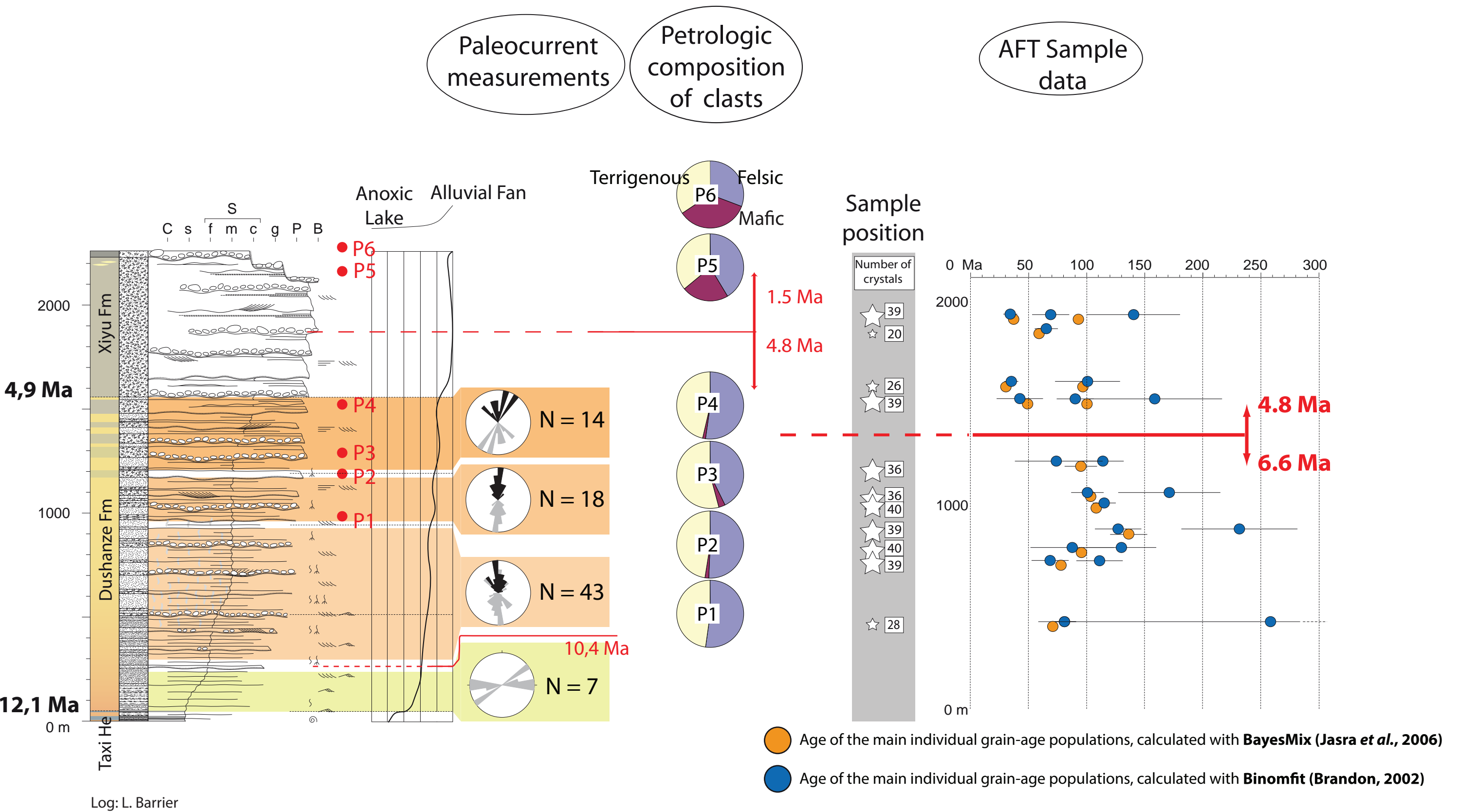
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Annexe - 1



Reconstructing the Late Palaeozoic – Mesozoic topographic evolution of the Chinese Tian Shan: available data and remaining uncertainties

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Abstract. The topographic evolution of continents and especially the growth and dismembering of mountain ranges plays a major role in the tectonic evolution of orogenic systems, as well as in regional or global climate changes. A large number of studies have concentrated on the description, quantification and dating of relief building in active mountain ranges. However, deciphering the topographic evolution of a continental area submitted to recurrent tectonic deformation over several hundred millions of years remains a challenge. Here we present a synthesis of the tectonic, geochronological and sedimentological data available on the intracontinental Tian Shan Range to describe its general topographic evolution from Late Palaeozoic to Early Tertiary. We show that this evolution has occurred in two very distinct geodynamic settings, initiating during the Carboniferous in an ocean subduction – continent collision tectonic context before becoming, from Early Permian, purely intra-continental. We show that during most of the Mesozoic, the topography is mostly characterized by a progressive general decrease of the relief. Nonetheless localized, recurrent deformation induced the formation of small-scale reliefs during that period. These deformations were driven by far field effects of possibly several geodynamic processes in a way that still remains to be fully understood.

1 Introduction

It is now largely accepted that the tectonic evolution of mountain ranges, climatic variations and feedback between the two is intimately linked to the topographic evolution of orogenic belts (e.g. Molnar and England, 1990; Batt and Braun, 1999; Whipple and Mead, 2006). Indeed, the topographic growth of an orogenic relief can influence the regional and sometimes global climate (e.g. Raymo and Ruddiman, 1992; Ramstein et al., 1997; Bookhagen and Burbank, 2006; Molnar et al., 2010). In turn, topographic changes due to climatic variations, which can modify the erosion rate and pattern in a range, can affect a tectonic evolution by, for example, impeding or accelerating the propagation of the deformation (e.g. Willett, 1999; Wobus et al., 2003; Thiede et al., 2004; Whipple, 2009). For those reasons, a large number of studies recently concentrated on describing the topographic evolution of active mountain ranges over a period of time spanning from a few thousands of years (e.g. Vassallo et al., 2007a; Nissen et al., 2009) to a few tens of million years (e.g. Metivier and Gaudemer, 1997; Sobel and Strecker, 2003; Yin, 2006; Liu-Zhen et al., 2008). In contrast, deciphering the topographic evolution of a mountain range area over several hundred millions of years remains a challenge and only a few studies have addressed that issue (e.g. Hendrix et al., 1992; Hendrix, 2000; Juez-Larré et al., 2010). However, understanding the long-term evolution of the topography is of primary interest to understand the relief and sedimentary basin dynamics in orogenic areas, the structural

and topographic inheritance in present day mountain ranges, or regional and global climate evolutions. The Tian Shan is one of the main intra-continental mountain ranges in Central Asia (Fig. 1). It forms a strong orographic barrier blocking the fresh and humid air masses coming from Siberia to the north and northwest, preventing them to reach the Tarim Basin and the Tibetan Plateau (e.g. Li et al., 2011; Zhao et al., 2013). The late Oligocene to Present growth of the Tian Shan Range resulted from crustal shortening related to the ongoing India-Asia collision (e.g. Tapponnier et al., 1986; Yin, 2010). However, the large-scale structure of the Tian Shan lithosphere was set by complex accretions of island arcs and amalgamation of continental blocks during the whole Palaeozoic (e.g. Windley et al., 1990; Charvet et al., 2011). This initial structure that was completely established by Late Permian, and especially the major tectonic boundaries separating the various terranes played a first order role in the subsequent evolution of the range.

The Palaeozoic and Cenozoic tectonic evolutions of the Tian Shan have been and are still widely studied and discussed. However, while the Mesozoic tectonic evolution of the range seems to have a major impact on its Tertiary development and topography (e.g. Dumitru et al., 2001; De Grave et al., 2007; Glorie et al., 2010; Jolivet et al., 2010), many uncertainties remain on the topographic evolution of the range during that period separating the two main orogenic events.

Here we present a short synthesis of the available data in order to draw a general picture of the topographic evolution of the Tian Shan from Late Palaeozoic to Early Cenozoic. We aim to describe the main phases of that topographic evolution, but also to point out the remaining inconsistencies between the various observations and to provide ideas for further investigations of the Mesozoic topography of Central Asia.

2 Synthesis of the available tectonic, geochronology, sedimentology and geomorphology data

A summary of the main data discussed below is presented in Fig. 2.

2.1 The Carboniferous – Permian evolution

It is commonly admitted that the Palaeozoic construction of the Tian Shan lithosphere through accretion of various continents and arc terranes ended during the Carboniferous by the collision between two blocks: the Yili – Central Tian Shan block and the North Tian Shan volcano-sedimentary complex – Junggar block (Figs. 1 and 3a) (e.g. Han et al., 2009; Wang et al., 2009; Charvet et al., 2011; Xia et al., 2012). However, the modalities of that last accretion event are still discussed. Two contrasting types of models are proposed.

The first models imply that the Late Carboniferous – Early Permian lavas that are widely exposed in the Yili block

were emplaced in a continental rift setting (e.g. Xiao et al., 1992; Che et al., 1996; Xia et al., 2004, 2008). Based on geochemical analysis and zircon U/Pb dating of the various Early to Middle Carboniferous volcanics and ophiolite suites occurring in the North Tian Shan block and indicative of an oceanic domain, Xia et al. (2005) suggested that this North Tian Shan Ocean was a Early Carboniferous short-lived ocean which closed before Late Carboniferous generating the North Tian Shan Middle Carboniferous ophiolite suite. This closure was followed by post-orogenic extension and transtension associated to continental rift-type volcanism in the Yili, Turfan and associated basins during the Late Carboniferous – Permian.

These models have been challenged by a number of tectonic, stratigraphic and geochemical evidences. Wang et al. (2009) indicated that magmatism in the Tian Shan has been continuous from Early Carboniferous to Permian with a magma composition evolving from essentially calc-alkaline and seldom transitional during the Carboniferous to coexisting calc-alkaline, transitional and alkaline during the Permian. Hence, alternative models proposed that the Carboniferous magmas would have been generated during southward subduction of the North Tian Shan Ocean (in these models a branch of the wide Paleo-Asian Ocean) underneath the Yili block (e.g. Wang et al., 2006, 2009; Han et al., 2009; Charvet et al., 2011). The North Tian Shan ophiolite sequence would correspond to an accretionary wedge developed in front of a Carboniferous forearc. This subduction ended with the Late Carboniferous – Early Permian closure of the North Tian Shan Ocean and the collision between the Yili block and the North Tian Shan – Junggar blocks (e.g. Gao et al., 1998; Wang et al., 2006; Charvet et al., 2007; Shu et al., 2010; Charvet et al., 2011) (Figs. 2 and 3a). The Permian magmas were then emplaced in a post-collisional setting and several authors suggested that slab break-off and lithospheric delamination may explain their mantle signature (Han et al., 1999, 2009; Chen and Jahn, 2004; Zhao et al., 2008; Shu et al., 2010). By Late Carboniferous – Early Permian, the Junggar Basin, disconnected from the Palaeo-Tethys ocean (Sha et al., 2011), initiated as a half-graben structure in a post-collisional extensional setting (Qiu et al., 2005, 2008; Yang et al., 2012).

All the models agree that during the Permian, the Tian Shan area was affected by a major transtension episode (Fig. 3b). Palaeomagnetic data analyses indicate that between the Upper Carboniferous and the Middle Permian, the Yili – West Junggar collage rotated counter-clockwise by $46.2^\circ \pm 15.1^\circ$ with respect to the Tarim and $31.6^\circ \pm 15.1^\circ$ with respect to Siberia (Wang et al., 2007b). As those continental blocks were amalgamated during the Upper Carboniferous this rotation has been accommodated by strike-slip motion along large lithospheric faults such as the North Tian Shan Fault, the Main Tian Shan Shear Zone, the Narat Fault or the Nikolaev Line in Tian Shan or the Erqishi shear zone in the Altay (Fig. 1). The total amount of E-W lateral displacement between Siberia and the Tarim is estimated at

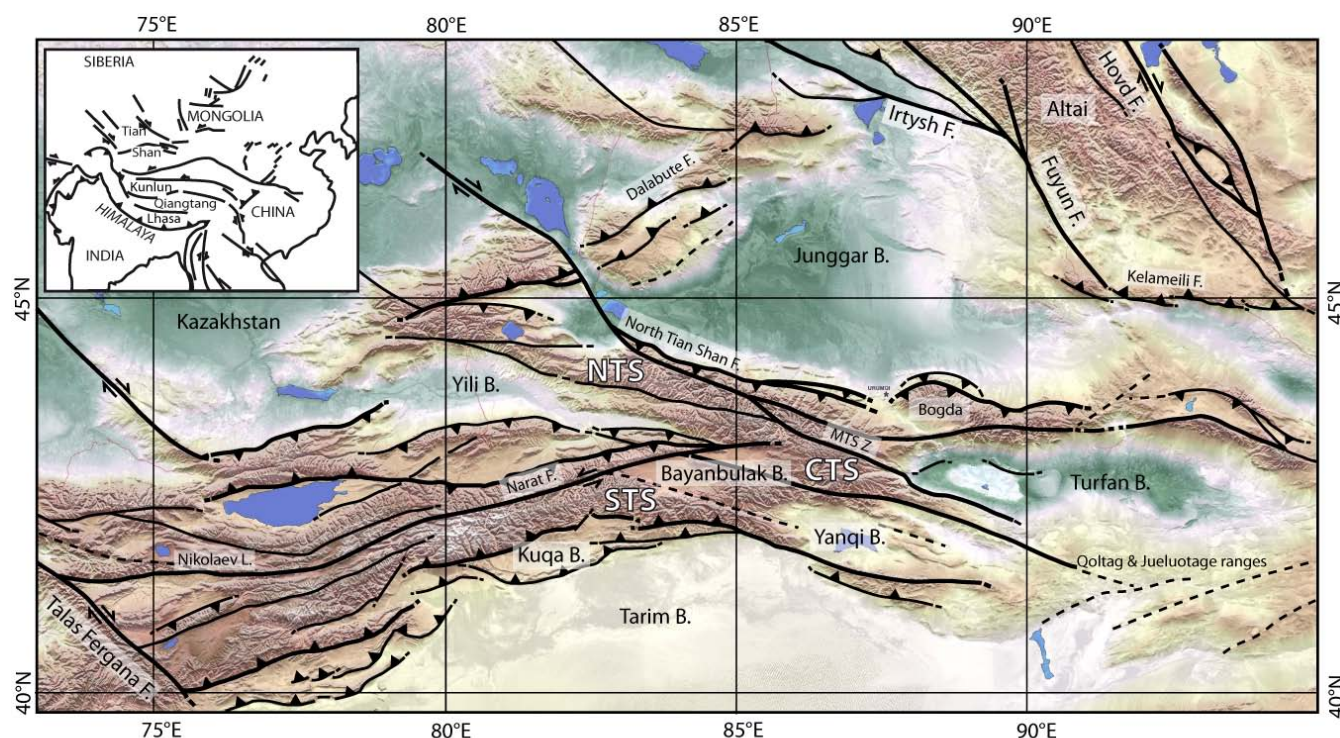


Fig. 1. General topographic and tectonic map of the Tian Shan Range and surrounding areas. MTS Z: Main Tian Shan Shear Zone; B. stands for basin; F. stands for fault. STS: South Tian Shan block. CTS: Central Tian Shan block. NTS: North Tian Shan block. The topography is a courtesy of S. Dominguez.

1160 ± 380 km (Wang et al., 2007b). Similar structures, like the Dalabute Fault to the west or the Kelameili Fault along the Altai, were also active around the Junggar Basin submitted to significant extension (e.g. Allen et al., 1991, 1995; Allen and Vincent, 1997; Qiu et al., 2005, 2008; Yang et al., 2012). Those large strike-slip faults have been active from Early Permian to Early Triassic (ca. 290–250 Ma; Laurent-Charvet et al., 2002; Shu et al., 2002; Wang et al., 2004, 2007a, de Jong et al., 2009) and partially controlled the emplacement of the Permian granitoids in Tian Shan (Wang et al., 2009). Zircon (U-Th)/He Permian ages indicate a strong cooling – exhumation phase within the Tian Shan Range at that time implying the build-up of a major positive topography affected by erosion (Jolivet et al., 2010) (Fig. 3b). Along the southern Junggar margin, detrital zircon U/Pb analysis indicate a major source area in the North Tian Shan block associated to a widening of the drainage pattern towards the Central Tian Shan block (Yang et al., 2012), again suggesting the presence of significant reliefs. Pull-apart basins formed along those strike-slip fault zones, sometimes associated to volcanism as along the Dalabute Fault (Allen and Vincent, 1997) (Fig. 1). De Jong et al. (2009) suggested that the continental breakup regime that allowed the emplacement of the Late Permian – Early Triassic volcanic traps in Siberia (Courtillot et al., 1999) could have generated a widespread extensive event within Central

Asia, reaching the Tian Shan area. South of the Bogda Shan and between the Tarim and South Tian Shan, extension allowed the formation of relatively small-scale but deep basins that connected with the close-by Palaeo-Tethys Ocean to the south (in the Tarim) and filled with Permian marine sediments (Li et al., 2005; Shu et al., 2010) (Fig. 3b). The occurrence of these sediments contrasts with the near absence of post-Carboniferous marine formations within the whole Altai – Junggar – Tian Shan area to the north that remained completely disconnected from any marine environment since the Early Permian. The Permian – Early Triassic closure of the Palaeo-Tethys and the subsequent Middle Triassic collision between the Qiangtang and Kunlun blocks to the south (Fig. 1) (e.g. Roger et al., 2008, 2010, 2011) might have ended that extension phase (e.g. Allen et al., 1991; Wartes et al., 2002; De Grave et al., 2007). In the Turfan area, sediments were then deposited in intra-mountain lacustrine and fluvial environments with a major sediment source from the Jueluotage Range to the south (Shao et al., 1999) (see Fig. 1 for location).

2.2 Triassic – Jurassic evolution

During the Early Triassic, the tectonic regime in the Tarim – Tian Shan – Junggar area changed to become clearly compressive (Fig. 2). Apatite fission track data from the Tian

	TIAN SHAN			JUNGGAR			TARIM / TURFAN			Topo.	Tecto.
	Tectonics	Geochrono.	Sedimento.	Tectonics	Geochrono.	Sedimento.	Tectonics	Geochrono.	Sedimento.		
65 Ma	CRETACEOUS	AFT and AHe data indicate localized strong uplift on inherited structures Uplift of Bogda Shan		Clockwise rotation Climax of basement highs Northward migration of the edge of the basin to near present-day position	Detrital zircon U/Pb suggest sediment recycling in piedmont	Unconformity		AFT data indicate exhumation in N Tarim	Detrital zircon U/Pb suggest uplift of basin margins in Tarim		Kohistan collision Lhasa collision
145 Ma		AFT data indicate isothermal stage throughout the whole range Detrital zircon U/Pb in Junggar Basin suggest uplift of the range									
200 Ma	JURASSIC	Transension along the N Tian Shan fault AFT and AHe indicate development of planation surface Cooling in Kyrgyzstan	Propagation of range front Erosion of sediment cover along S and N piedmonts	Thermal relaxation remains main subsidence mechanism S to N propagation of basement highs Clockwise rotation of Junggar Both normal and reverse faults around the basin	Detrital zircon U/Pb suggest sediment recycling in piedmont Volcanic ash deposits	Aridification Kalaza conglomerates Onset of progradation phase suggesting relief building in Tian Shan Lacustrine deposits Restricted brackish-water sediments in south Junggar	Extension in Turfan and Tarim	Detrital zircon U/Pb suggest sediment recycling in piedmont	Kalaza conglomerates Uplift of Bogda Shan suggested by alluvial fans Sediments from S and N in Turfan		
251 Ma	TRIASSIC	AFT indicate continuous cooling	Onset of coal-bearing sedimentation	Thrusting along the edges of the basin End of tectonic subsidence Growth of basement highs (compression) Onset of compression	Detrital zircon U/Pb indicate strong widening of drainage towards the Tian Shan	No obvious unconformity Onset of coal-bearing sedimentation		Compressive basement highs in Turfan Extension in Turfan with strong thermal subsidence Onset of compression in N Tarim	Onset of coal-bearing sedimentation N Tarim: sediment source in South and Central Tian Shan Conglomerates		
299 Ma	PERMIAN	Major transension (1 160 +/- 380 km accommodated) and formation of lithospheric strike-slip faults	ZHe ages indicate strong exhumation Post-collisional granitoids	Growth of basement highs (extension) Significant extension. Rotation. Active strike slip faults to the E and W Half grabben basin		Widening of the drainage pattern Major source in North Tian Shan Clastic sediments. No connexion with marine areas	Extension. Localised, deep basins		Marine sediments. Connexion with Palaeo-Tethys Ocean		CAOB accretion
	Carboniferous	Accretion of continental blocks and arc terranes. Modalities discussed		Collision with North Tian Shan.							

Fig. 2. Summary of the main Mesozoic tectonic, geochronological and sedimentary constraints in the Tian Shan Range and the Junggar and Tarim-Turfan basins. The data indicated in red are apparently in contradiction with the interpretation derived from other available data at the same time. The data in blue is contradicting the others but can be explained by local mechanisms. Column Topo. indicates the relative increase (wide black line) or decrease (thin black line) of the Tian Shan general topography. Column Tecto. recalls the major geodynamic events that affected Central Asia.

Shan clearly indicate a continuous cooling phase during the Triassic and Jurassic, even if the cooling rate become extremely slow during Middle to Late Jurassic (e.g. Dumitru et al., 2001; De Grave et al., 2007; Jolivet et al., 2010). Basement highs (imaged by seismic lines) that already started to uplift during the Carboniferous compressive (He et al., 2005) and Permian transpressive event in the western Junggar Basin continued to grow during the Triassic (Liu et al., 2006). However, within the Turfan Basin, tectonic extension still prevailed during the Early Triassic, associated to a strong thermal subsidence (Shao et al., 1999). Sediments were still derived from the south. In the northern Tarim Basin, south to southeastward directed palaeocurrents, as well as detrital zircon U/Pb dating, indicate a sediment source in the Central Tian Shan block and along the southern margin of the South Tian Shan block (Li et al., 2004, Li and Peng, 2010). This, associated with the occurrence of Lower Triassic conglomerates in the Early Triassic Ehuobulake Formation of the Kuqa sub-basin (the names of the various sedimentary formations can be found for example in Hendrix et al., 1992, Shao et

al., 1999; Jin et al., 2008 and Qiu et al., 2008), is consistent with the compression-related exhumation phase recognized by the low temperature thermochronology data within the range (Dumitru et al., 2001; Li et al., 2004; Jolivet et al., 2010). Detrital zircon U/Pb dating in the Middle to Late Triassic along the southern edge of the Junggar Basin (Xiao-quangou Group) indicate a strong widening of the drainage system towards the Central Tian Shan block associated to the progressive erosion of the Permian – Early Triassic topography (Yang et al., 2012). From Middle Triassic, basement highs also started to develop in the Turfan Basin, probably in a compressional setting (Shao et al., 1999). During the same period, tectonic subsidence ceased in the Junggar Basin, which became a slowly subsiding depression (Yang et al., 2012).

Except in the northern and western edges of the Junggar Basin, where Late Triassic thrusting occurred (Allen et al., 1991; Liu et al., 2006), and in the south Junggar, where small reverse movements took place (Yang et al., 2012), there is no obvious discordancy between the Upper Triassic and the

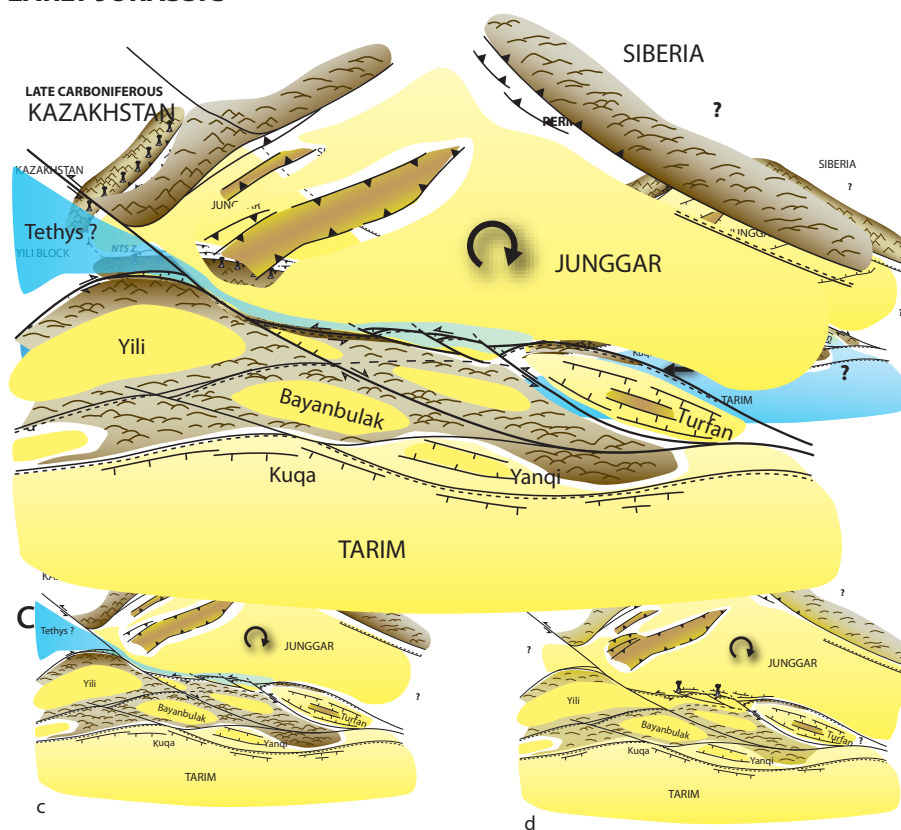
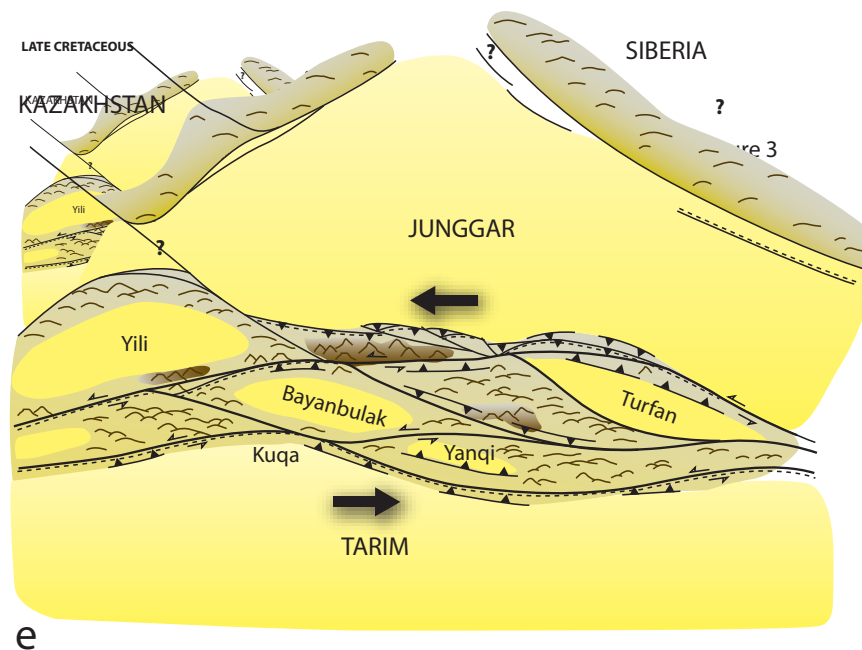
EARLY JURASSIC**LATE CRETACEOUS**

Fig. 3. Reconstitution of the main topographic pattern of the Tian Shan area through time. The brown areas correspond to mountains, the yellow areas to continental basins and the blue areas to marine basins. On Fig. 2a, NTS Z: North Tian Shan Shear Zone; CTS Z: Central Tian Shan Shear Zone; STS Z: South Tian Shan Shear Zone. See text for complete discussion.

Lower Jurassic series (e.g. Sha et al., 2011). Indeed, from the Middle Triassic to the Middle Jurassic, a widespread coal-bearing sediment sequence developed throughout the Tian Shan and surrounding basins, most probably in a warm humid climate (Hendrix et al., 1992; Pfretzschner et al., 2005; Li et al., 2004; Tükten et al., 2004; Ashraf et al., 2010; Pan et al., 2013).

During the same period, the relative topography between the range and the northern margin of the Tarim decreased while the Junggar Basin widened towards the south (Li et al., 2004; Fang et al., 2005; Li and Peng, 2010; Yang et al., 2012). In Early Jurassic (Sinemurian Badaowan Fm.), a brackish-water fauna and flora (*Waagenoperma* bivalve assemblages and *Tasmanites* sp. alga) developed in some restricted areas along the southern margin of the Junggar Basin (Sha et al., 2011). Pan et al. (2013) suggested that small short-lived transtensional basins formed through reactivation of the North Tian Shan Fault. A contemporaneous increase in the water level of the North Tethys Ocean allowed a transgression event from the west along the southern margin of the Junggar (Fig. 3c). Small-scale tectonic activity, driven by a Jurassic – Early Cretaceous clockwise rotation of the Junggar block (Hou et al., 2009) is also observed in the northern and western Junggar during the Early and Middle Jurassic, both on thrust faults (Allen and Vincent, 1997) and normal faults (Liu et al., 2006). Seismic data interpretation indicates that this tectonic activity induced the uplift of topographic highs in the basin that propagated from south to north. However, the main driving mechanism for the subsidence of the Junggar basin would remain the thermal relaxation (Hou et al., 2009).

During the Early Jurassic (Badaowan Fm.), the Turfan Basin was also affected by extension that controlled its sedimentation during most of the Jurassic (Fig. 2). The sediments were still deposited in an environment influenced by surrounding topographic highs. Their main source area remained in the south (Qoltag and Jueluotage Ranges, Fig. 1), but small alluvial fans also formed on the northern edge of the basin, along the Bogda Shan. However, the topography of that range remained low (Shao et al., 1999, 2003). Normal faulting also affected the northern Tarim margin during the Jurassic (Allen et al., 1991).

By Middle Jurassic, a widespread planation surface started to develop over most of Central Asia, extending probably from Northern Tibet to SE Siberia (e.g. Allen et al., 2001; Vassallo et al., 2007b; Jolivet et al., 2007, 2009, 2010, 2013a). Remnants of that surface are still preserved in the central and northern part of the Tian Shan Range (e.g. Jolivet et al., 2010) (Fig. 3d). Over the same period, the apatite fission track data obtained from the central Chinese Tian Shan show a slow erosion period (isothermal to very slow cooling pattern compatible with a tectonically quiet phase) throughout the whole range (Dumitru et al., 2001; Jolivet et al., 2010; Glorie et al., 2010). However, further west in Kyrgyzstan, De Grave et al. (2007) indicate continuous cooling from

180 to 110 Ma. In the Junggar Basin, the long-lasting general retrogradation phase that initiated in the Late Carboniferous ended with the lacustrine facies of the Middle Jurassic Xishanyao Formation. The late Middle Jurassic change to a progradation phase suggests a renewed topography building in the Tian Shan or a strong climate change (Yang et al., 2012). However, the warm humid climate conditions that prevailed since Middle Triassic did not change until the onset of a strong aridification during the Late Jurassic (Qigu Formation) (Allen et al., 1991; Hendrix et al., 1992; Parrish, 1993; Shao et al., 2003; Li et al., 2004; Pan et al., 2013). Detrital zircons U/Pb dating still suggest a wide drainage system encompassing the North Tian Shan and the Central Tian Shan blocks but local recycling of the Early Mesozoic cover may have also occurred (Yang et al., 2012). The Middle and Upper Jurassic series of the southern Junggar are also marked by the occurrence of widely distributed volcanic ashes of intra-continental basaltic composition interbedded within the sediments (Wang and Gao, 2012; Yang et al., 2012).

As previously said, the Late Jurassic period is marked by a strong aridification and the widespread occurrence of the conglomeratic Kalaza Formation in the Junggar, northern Tarim and Turfan basins (Fig. 2). In the south Junggar and north Tarim, conglomerates are derived from the Tian Shan Range while in the Turfan Basin, southward thinning of the Kalaza Formation from the northern edge of the basin indicates a source in the Bogda Shan then forming a topographic high (e.g. Hendrix et al., 1992, 2000; Shao et al., 1999; Sha et al., 2011). Detrital zircons U/Pb dating in the southern Junggar indicates, that recycling of the Mesozoic cover sequence associated to some topographic growth initiated during Middle Jurassic (Yang et al., 2012). In the Kuqa sub-basin, similar detrital zircons U/Pb dating indicates a sediment source in the South Tian Shan where the sediment cover was progressively eroded. This front of the range progressively migrated towards the south, increasing the difference in topography between the basin and the uplifting range (Li et al., 2004).

Finally, at least in the Junggar Basin, the Lower Cretaceous series are described as unconformably lying on the Kalaza Formation (Sha et al., 2011), this again indicating a Late Jurassic tectonic activity (e.g. Hendrix et al., 1992, 2000; Shao et al., 1999; Sha et al., 2011). However, these tectonic movements are not or only poorly recorded by the low temperature thermochronology data in the central part of the Chinese Tian Shan (Dumitru et al., 2001; Jolivet et al., 2010). Contemporaneous cooling is only observed in the Kyrgyz Tian Shan further west (e.g. De Grave et al., 2007). Consequently, this near absence of cooling evidences in the low temperature thermochronology record, like the formation and the extensive preservation of the Jurassic planation surface within the range, are difficult to reconcile with Middle to Late Jurassic topographic growth and erosion revealed by the sedimentological and detrital zircon U/Pb data.

2.3 Cretaceous – Early Palaeogene evolution

Sediment analysis and detrital zircon U/Pb dating indicate that tectonic uplift continued during the Early Cretaceous along the southern edge of the Junggar Basin (Fang et al., 2006; Yang et al., 2012). An increase in the amount of detrital zircons derived from the North Tian Shan block suggests a noticeable uplift and denudation of this block. This uplift potentially formed a barrier to the sediments coming from further south. The southern edge of the basin was slightly uplifted and migrated northward to approximately its present-day position (Fang et al., 2006). The clockwise rotation of the Junggar block continued and intra-basin highs, which initiated during the Jurassic, reached their climax in Early Cretaceous with relative topographies up to 600 m based on erosion surfaces imaged by seismic data (Hou et al., 2009). A similar basin margin uplift is observed in the north Tarim Basin margin. Indeed, detrital zircon U/Pb analysis indicate an increase in denudation of the Central Tian Shan and South Tian Shan blocks associated to an increase of the differential topography between the basin and the range (Li and Peng, 2010).

Several authors suggested that the Early Cretaceous deformation around the Tian Shan Range resulted from the collision of the Lhasa block to the south (Figs. 1 and 2) (e.g. Hendrix et al., 1992; Gu, 1996; Fang et al., 2006a; De Grave et al., 2007). However, the absence of Late Jurassic – Early Cretaceous cooling episode in the Songpan – Garze area (East Tibet), immediately north of the Bangong – Nujiang suture Zone between the Lhasa and Qiangtang blocks suggests that this accretion generated only a very limited deformation (Coward et al., 1988; Roger et al., 2008, 2010, 2011). Nonetheless, further to the north, between the Kunlun Ranges and the Tarim Basin, low temperature thermochronology and sediment analysis do indicate some small-scale Cretaceous vertical movements potentially linked to the Lhasa collision (e.g. Vincent and Allen, 1999; Jolivet et al., 1999, 2001; Marshall et al., 2000). Within the Chinese part of the Tian Shan Range, the low temperature thermochronology data indicate that the near isothermal stage that started during the Middle Jurassic still prevailed during the Cretaceous (e.g. Jolivet et al., 2010). In Kyrgyzstan, this phase is also recorded from late Early Cretaceous up to the Palaeogene (e.g. De Grave et al., 2007; Glorie et al., 2010). As for the Middle and Late Jurassic, this contradicts the sedimentological (occurrence of conglomerates in the Cretaceous series) as well as recycling observed in the detrital zircon U/Pb data. However, Zhang et al. (2009), using apatite fission track analysis, reported evidences of Late Cretaceous exhumation near Aksu, along the northern margin of the Tarim Basin.

Eventually, the late Late Cretaceous – Early Palaeogene marks the onset of a new deformation phase that will culminate with the ongoing orogeny (Figs. 2 and 3e). Within the range, the low temperature thermochronology data show that

localized and rapid deformation and uplift occurred as early as 65 to 60 Ma on large inherited structures such as the Nikolaev line (Jolivet et al., 2010). Apatite fission track analysis in the Bogda Shan Range also show an onset of cooling and uplift around 65 Ma (Wang et al., 2008). The geodynamic mechanism driving those localized, Early Cenozoic deformations is still poorly known and might correspond either to far-field effects of the closure of the Mongol-Okhotsk Ocean in SE Siberia (Wang et al., 2008; Jolivet et al., 2009, 2010) or to terranes accretions such as the Kohistan-Dras Arc or the Kabul Block to the west in Afghanistan and northwestern Pakistan (Searle, 1991; Hendrix et al., 1992; Jolivet et al., 2010).

3 Discussion

Many evidences – tectonic, geochemical, geochronological and sedimentological – are in favour of the Late Carboniferous subduction – collision model (e.g. Gao et al., 1998; Wang et al., 2006; Charvet et al., 2007; Shu et al., 2010; Charvet et al., 2011). While the alternative rift model (e.g. Xiao et al., 1992; Che et al., 1996; Xia et al., 2004, 2008) would have generated a relatively low-relief topography distributed over most of the present day Chinese Tian Shan area, the subduction – collision model probably gave birth to a dissected, high relief, Andean- to Alpine-type topography (Fig. 3a). This kind of topography is consistent with the Late Palaeozoic – Early Mesozoic erosion of metamorphic basement rocks (e.g. Yang et al., 2012).

The Permian transtension event and the formation of large pull-apart basins along major strike-slip faults was certainly one of the main tectonic events in the geodynamic history of the range. The Late Carboniferous – Early Permian collision-related topography was dissected as strike-slip motion created the Yili, Bayanbulak, and Turfan basins. This event set up the tectonic framework of the Mesozoic and Tertiary Tian Shan (Fig. 3b). Continuous tectonic and/or thermal subsidence prevailed in the Junggar Basin during all that period. It was possibly linked to the general extensional setting that affected the northern part of Central Asia (De Jong et al., 2009), and allowed the deposition and preservation of the sediments issued from the widespread denudation of the adjacent areas as recorded by the low temperature thermochronology data (e.g. Dumitru et al., 2001; Jolivet et al., 2010).

The onset of renewed compression marked by basin inversion and basement cooling during the Early Triassic marks the end of the strike-slip phase (e.g. Dumitru et al., 2001; Liu et al., 2006; Jolivet et al., 2010). This Early Triassic event, probably linked to the final closure of the Palaeo-Tethys Ocean to the south (e.g. Roger et al., 2008, 2010, 2011), was followed by a general decrease in relief and the progressive formation of the flat topography that will characterize Central Asia during the Mesozoic. Local extension such as in the

Turfan Basin and thermal subsidence in the Junggar Basin maintained active sedimentation areas filled with sediments derived from erosion of remaining topographic highs.

Except for the development of basement highs within the Junggar Basin, flattening of the topography further continued during the Early and Middle Jurassic. The development of a flat topography is consistent with the widespread occurrence of a coal system throughout the whole area (e.g. Sha et al., 2011) (Figs. 3c and 2d). The relief between the Tian Shan area and the surrounding basins probably decreased through erosion. This phase of slow erosion is clearly recorded by the low temperature data that show a protracted isothermal stage during the whole Jurassic period (e.g. Dumitru et al., 2001; Jolivet et al., 2010; Glorie et al., 2010). However, tectonic activity probably still occurred along some of the major Permian tectonic structures, creating reliefs. For example, the formation of small-scale extensional (probably pull-apart) basins along the southern edge of the Junggar Basin, associated with volcanism, suggest strike-slip faulting along the North Tian Shan Fault zone (Pan et al., 2013). Evidence for local relief building is further attested by the observed, progressive recycling of the Mesozoic sediments along the southern margin of the Junggar (Yang et al., 2012). However, topographic highs were certainly not important and the peneplanation phase probably reached its maximum during the Middle or Late Jurassic like in most of Central Asia (e.g. Vassallo et al., 2007b; Jolivet et al., 2001, 2007).

The occurrence of the major alluvial fan system of the Late Jurassic Kalaza Formation seems to mark the onset of a new tectonic and topographic phase. While the low temperature thermochronology data obtained on basement rocks in the Chinese Tian Shan still indicate slow, erosion-driven cooling (Dumitru et al., 2001; Jolivet et al., 2010), sedimentological and tectonic evidences imply tectonic changes of topography. The Late Jurassic period is also marked by aridification (Allen et al., 1991; Hendrix et al., 1992; Parrish, 1993; Shao et al., 2003; Pan et al., 2013). The formation of the Kalaza alluvial fan could be linked to this climate change but relief building is also attested by the unconformity between the Late Jurassic and Lower Cretaceous sediments in the Tarim and Junggar basins (e.g. Hendrix et al., 1992, 2000; Shao et al., 1999; Sha et al., 2011). Similarly, Late Jurassic – Early Cretaceous basement cooling observed in the Kyrgyz Tian Shan suggest tectonic activity (De Grave et al., 2007). Altogether this tectonic phase remains poorly understood. Once again, relief building during that period must have been limited both in magnitude and extend. Widespread formation of relief is not compatible with the preservation – up to now – of the Mesozoic planation surface within the range. It has been suggested that the deformation observed in the Tian Shan was linked to the Early Cretaceous collision between the Lhasa block and the southern margin of Eurasia (e.g. Hendrix et al., 1992; Gu, 1996; Fang et al., 2006; De Grave et al., 2007). However, if the Late Jurassic Kalaza Formation marks the onset of the tectonic phase, it initiated prior to the colli-

sion. Furthermore this collision generated only a very limited amount of deformation and exhumation along the southern margin of the Qiangtang block (e.g. Coward et al., 1991; Roger et al., 2008, 2010, 2011). Why then did this collision affect the Central Asian crust several thousands of kilometers to the north? Another major geodynamic event, the closure of the Mongol-Okhotsk Ocean, affected Central Asia during the Mesozoic (e.g. Zonenshain et al., 1990a, b; Nie et al., 1990; Zorin et al., 1990; Metelkin et al., 2007, 2010). While the age of the final closure is still largely discussed (see Jolivet et al., 2009, 2013b and references therein), it probably occurred during the Cretaceous in southern Siberia. The far-field effects of that oceanic closure (combined or not with the effects of the Lhasa collision) could also be the driving mechanism for the Late Mesozoic deformations in the Tian Shan area.

Finally, the Late Cretaceous – Early Palaeogene reactivation of the Palaeozoic (mainly Permian) major strike-slip structures initiated the building of the present day Tian Shan topography (Fig. 3e). The relief first formed in highly localized zones along strike-slip faults like the Narat Fault (Jolivet et al., 2010). The preservation of large remnants of the Mesozoic planation surface implies that exhumation and erosion within the present day Tian Shan is still low and remains localized along discrete inherited tectonic structures.

4 Conclusions

The Late Palaeozoic and Mesozoic topographic evolution of the Tian Shan Range has been controlled by two, very different mechanisms:

- During the Late Carboniferous – Early Permian, oceans are still present around the Palaeo-Tian Shan and the topography is controlled by subduction and collision events. This probably first generated an Andean-type topography followed by an Alpine-type topography after the final closure of the Palaeo-Asian Ocean.
- After this final oceanic closure and up to now, the Tian Shan became an intra-continental area, only affected by the far field effects of continental collisions and oceanic closures that happened several thousands of kilometers away. During this second phase, the Middle-Permian – Early Palaeogene topographic evolution has been marked by the Middle to Late Permian dislocation of the Late Carboniferous relief and its progressive planation.

Indeed, several small-scale tectonic deformations affected the Tian Shan area during the Mesozoic. However the associated topographic changes remained limited and probably localized. While the Cimmerian collisions certainly induced the Triassic compressive event that ended the major

Permian strike-slip movements, the Late Jurassic and Cretaceous phase is still poorly understood. If the Cretaceous collision of the Lhasa block might have generated some of the stress that induced Late Mesozoic deformation in the Tian Shan, the effects of the Mesozoic closure of the Mongol–Okhotsk ocean should also be considered.

Finally, the synthesis of all the available data, and especially the discrepancy between the low temperature thermochronology data and the sedimentology data during the Late Mesozoic period demonstrates the need of combining both methods to decipher in details the topographic evolution of a region.

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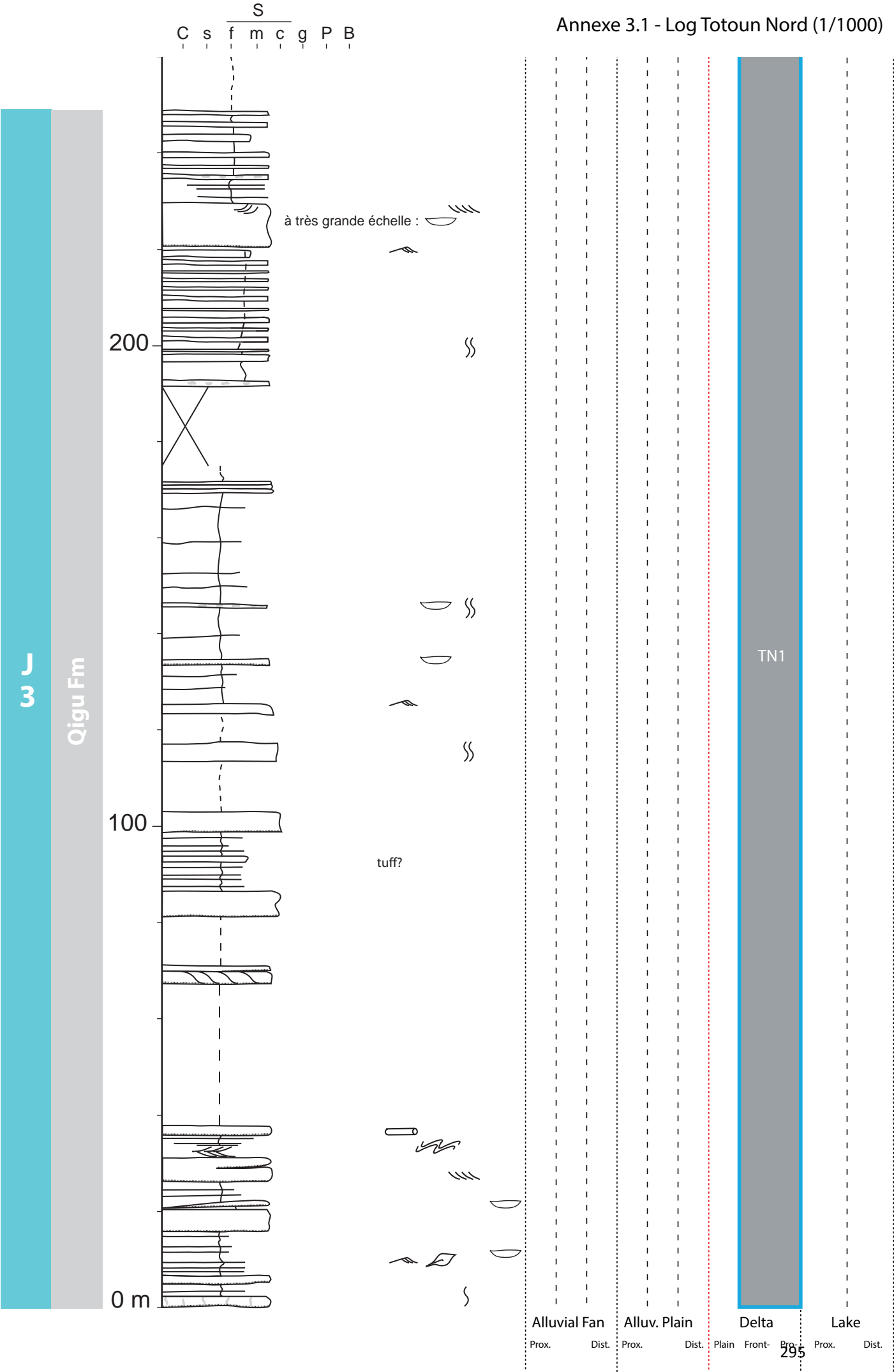
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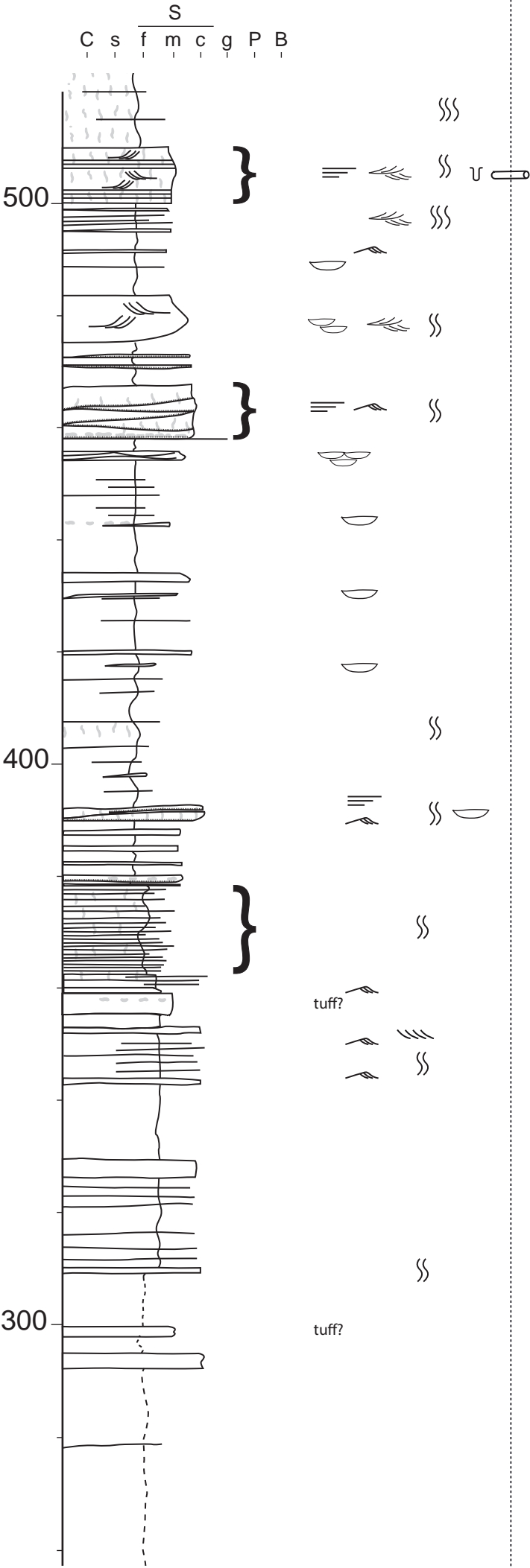
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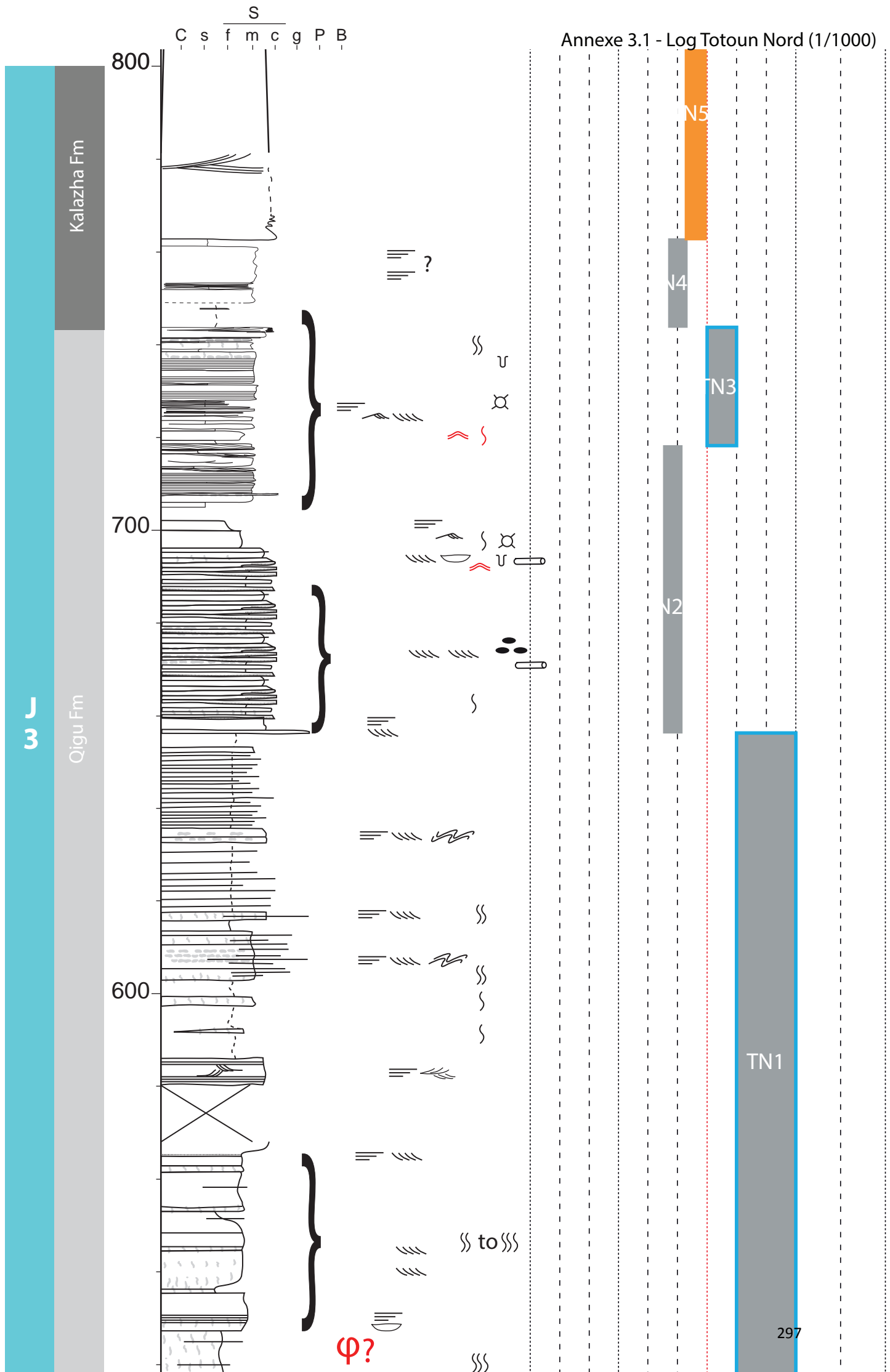
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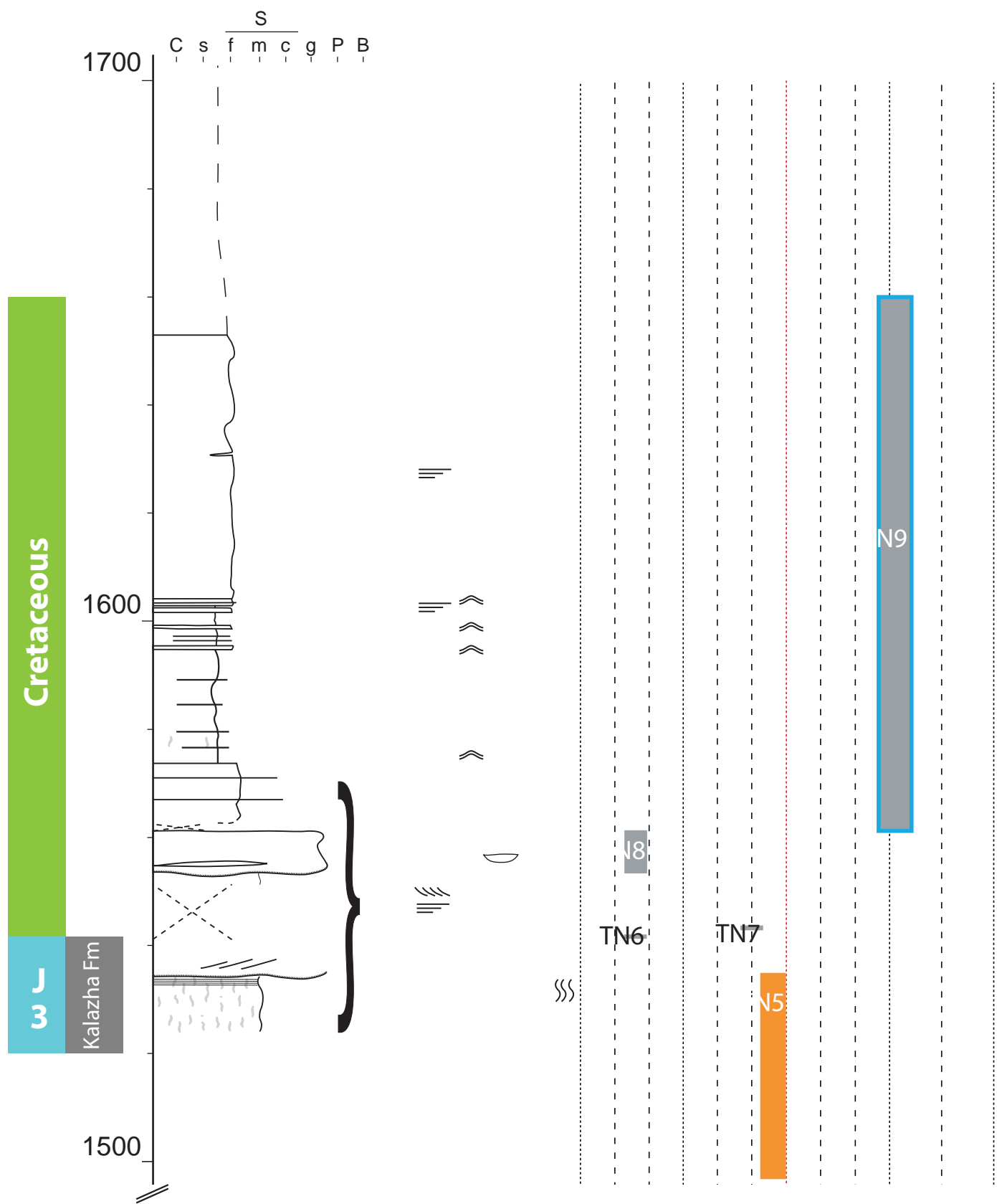
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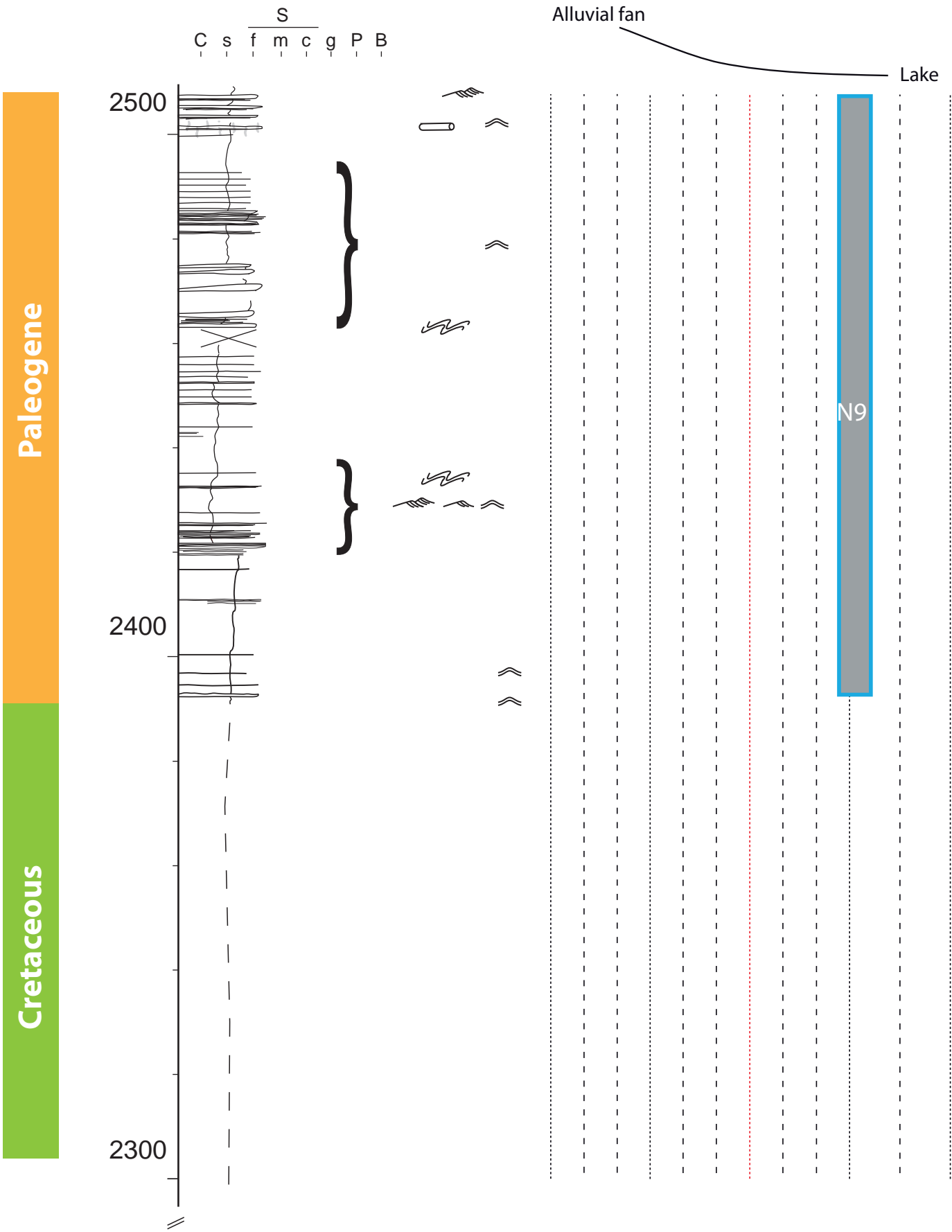




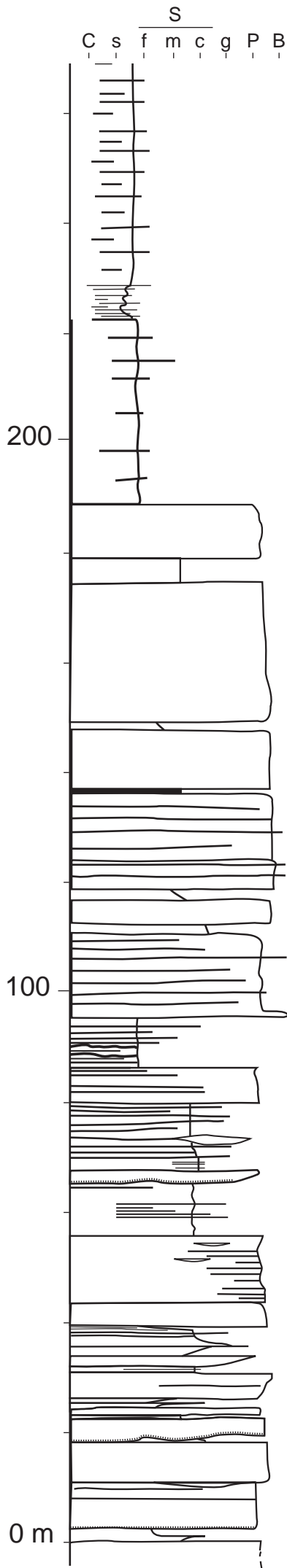
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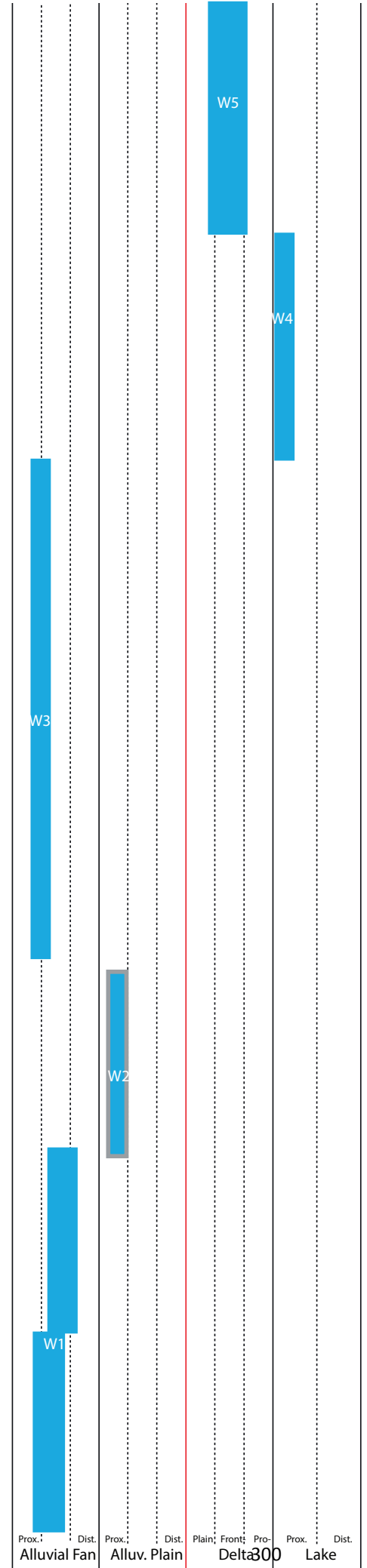


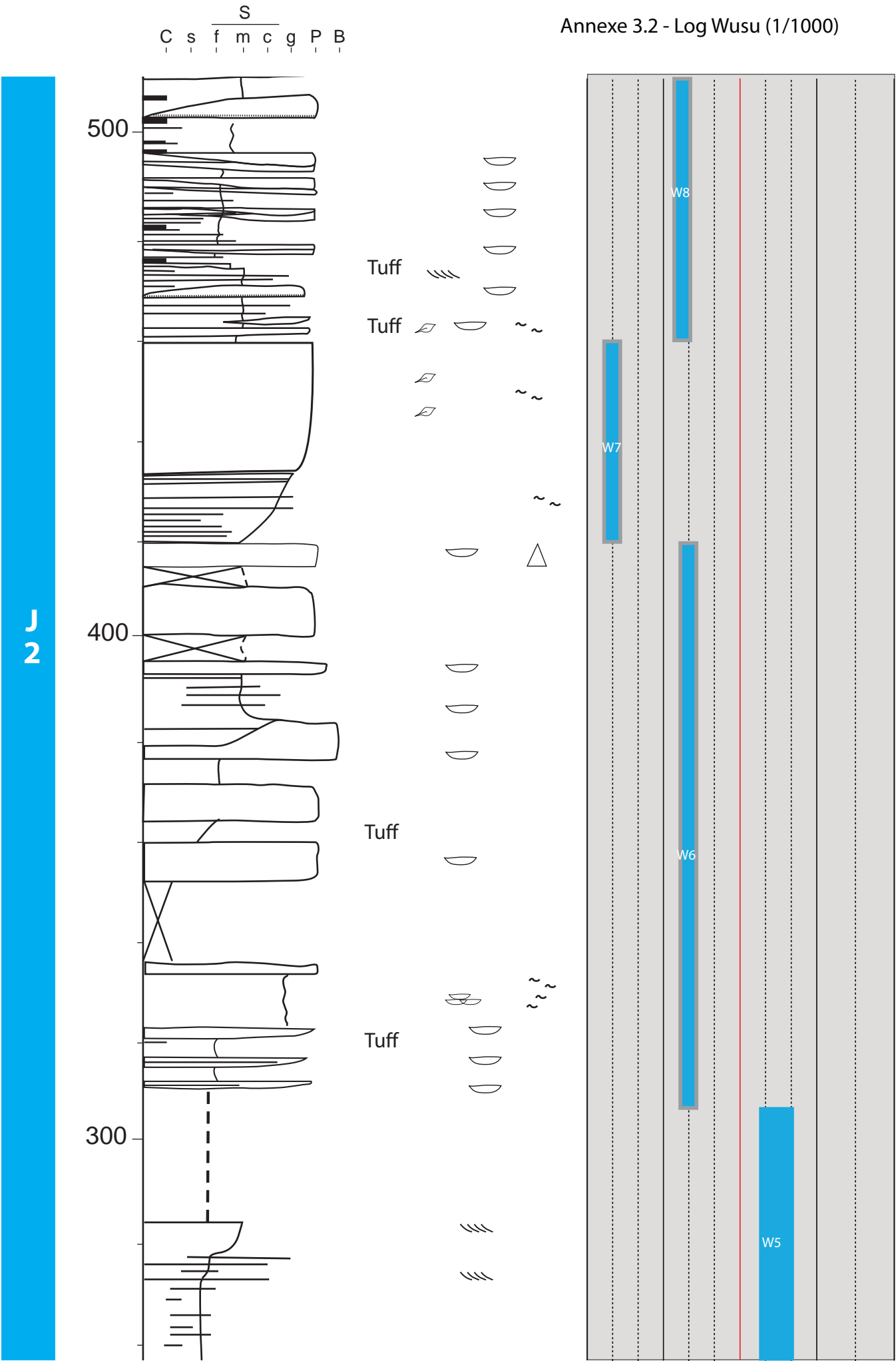


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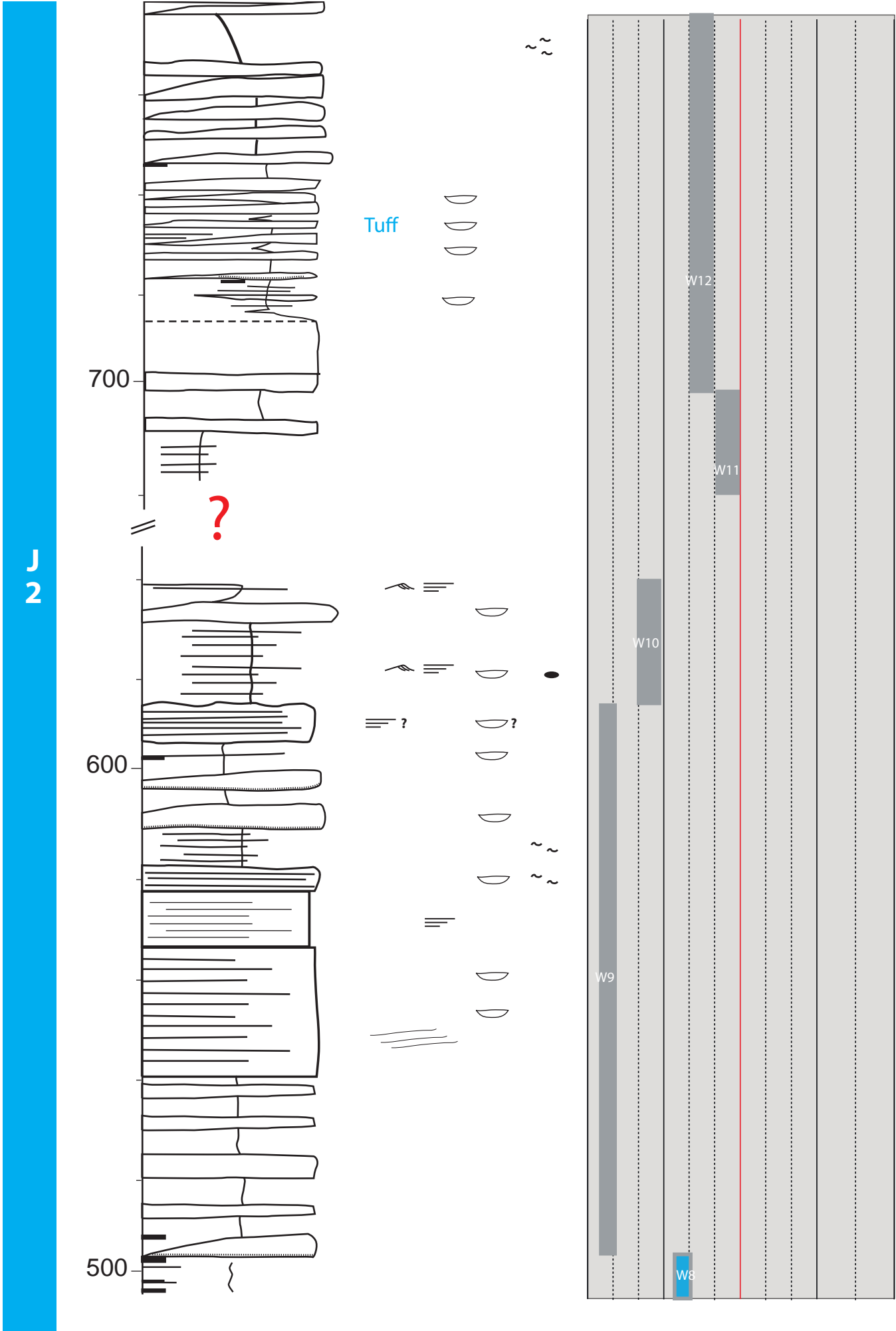


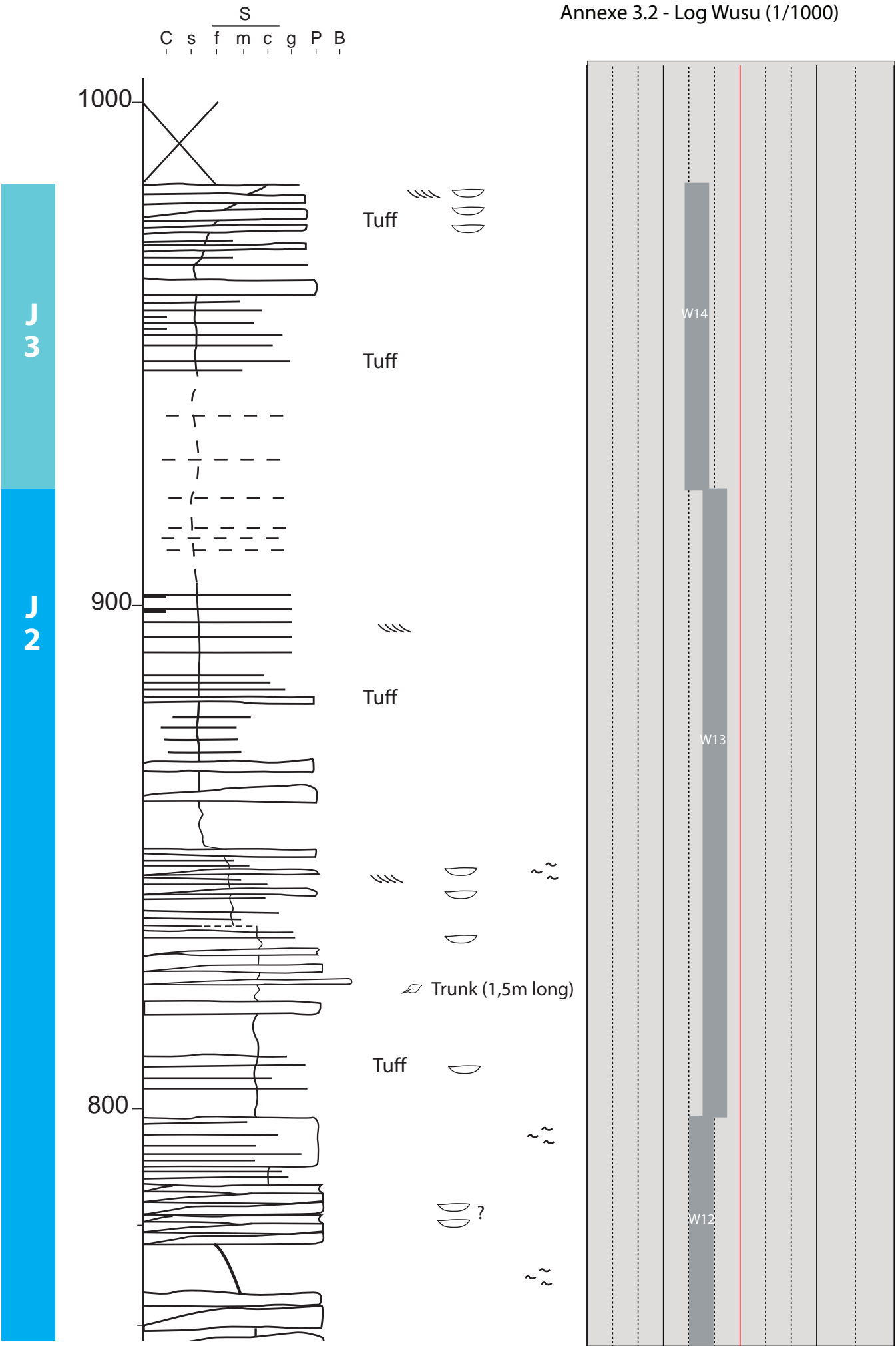
Annexe 3.2 - Log Wusu (1/1000)



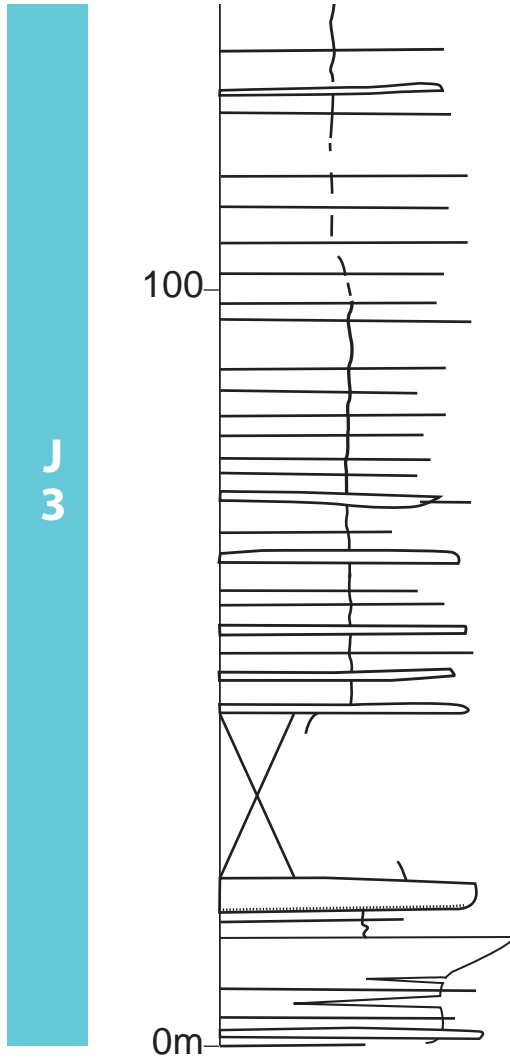


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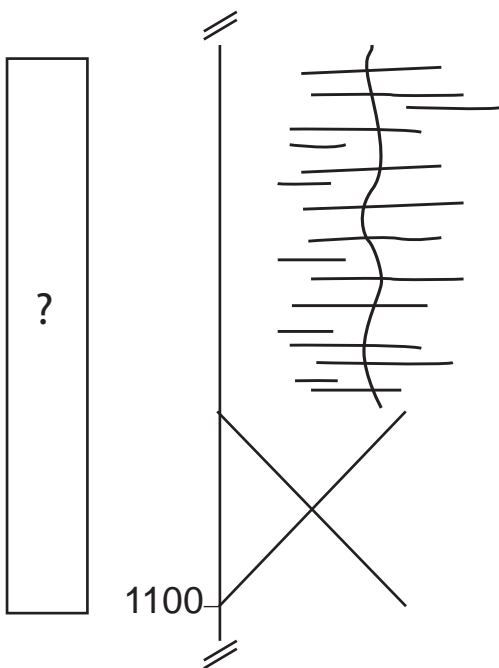
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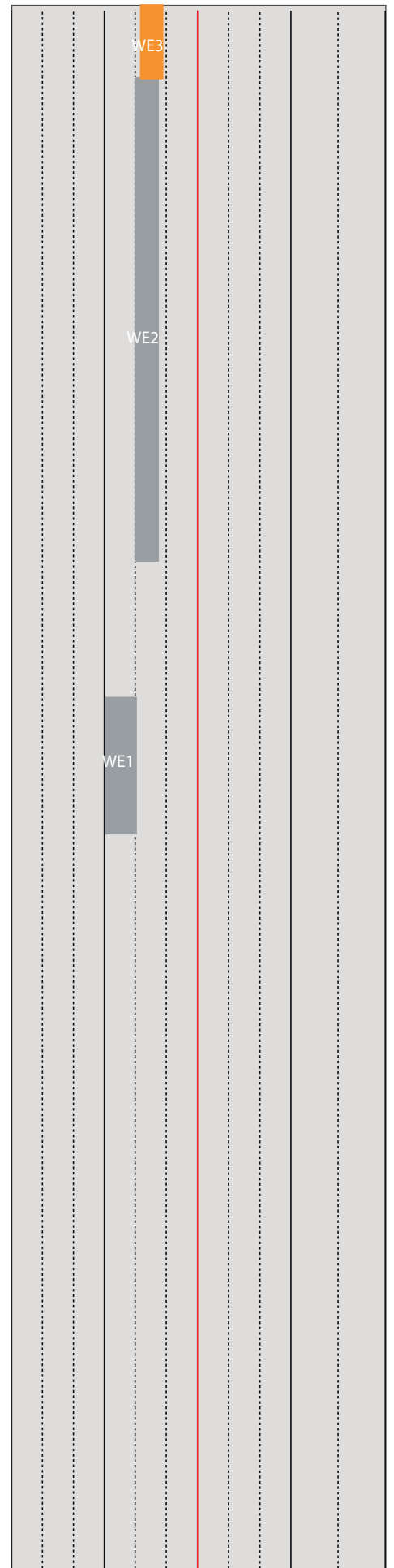
λ?

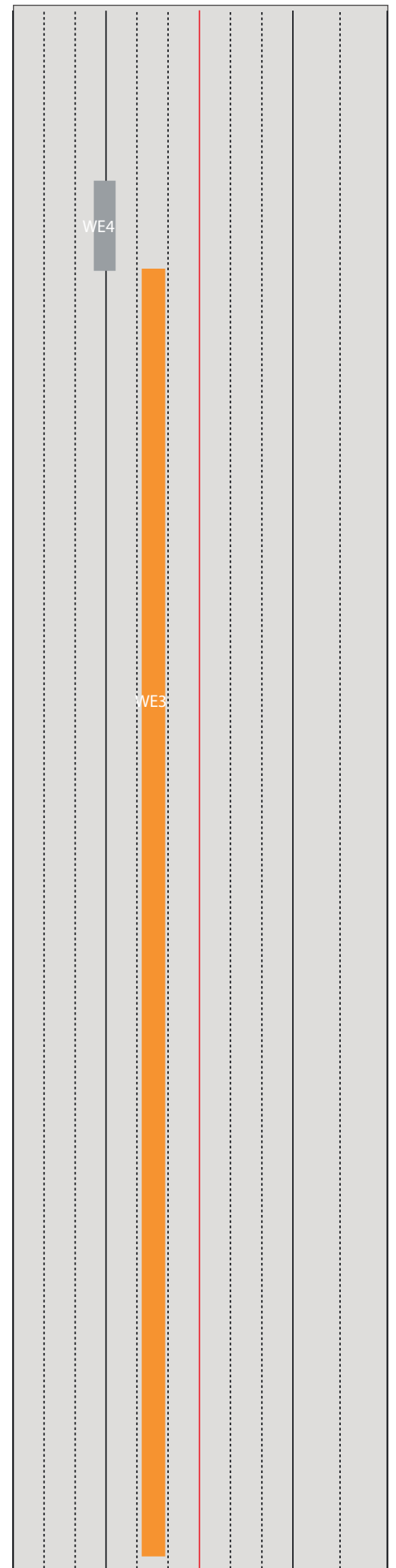
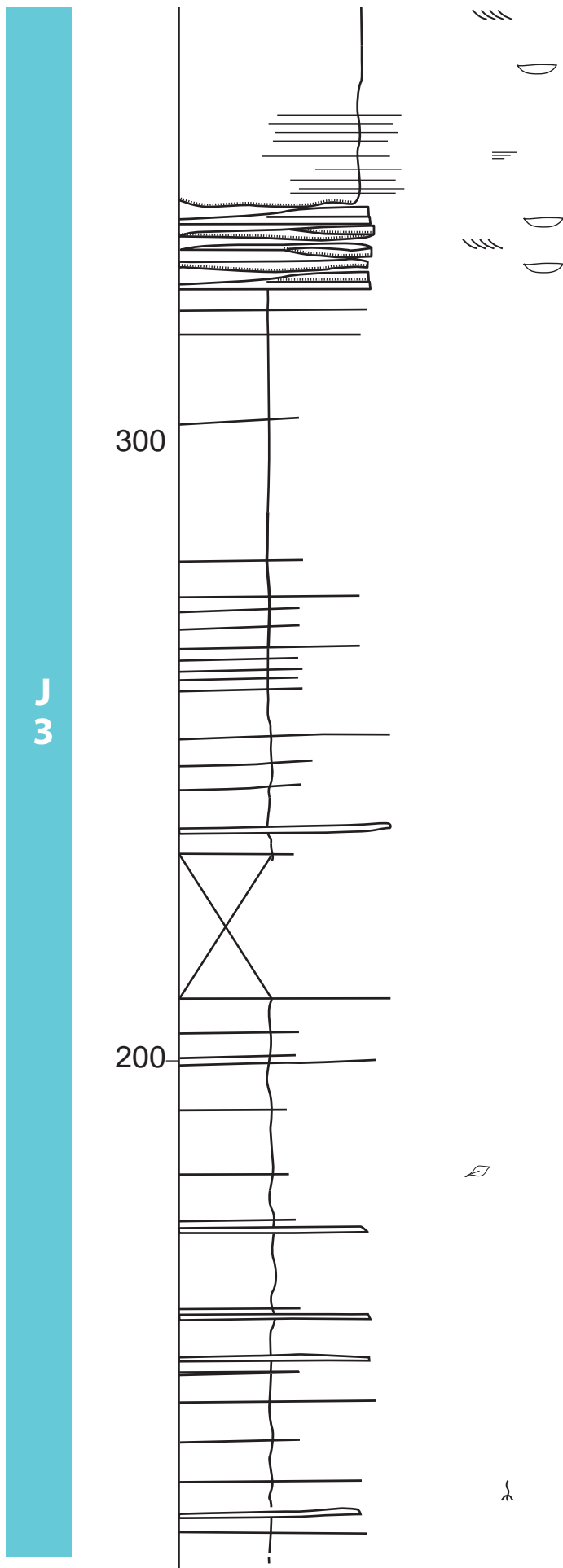
GPS W100

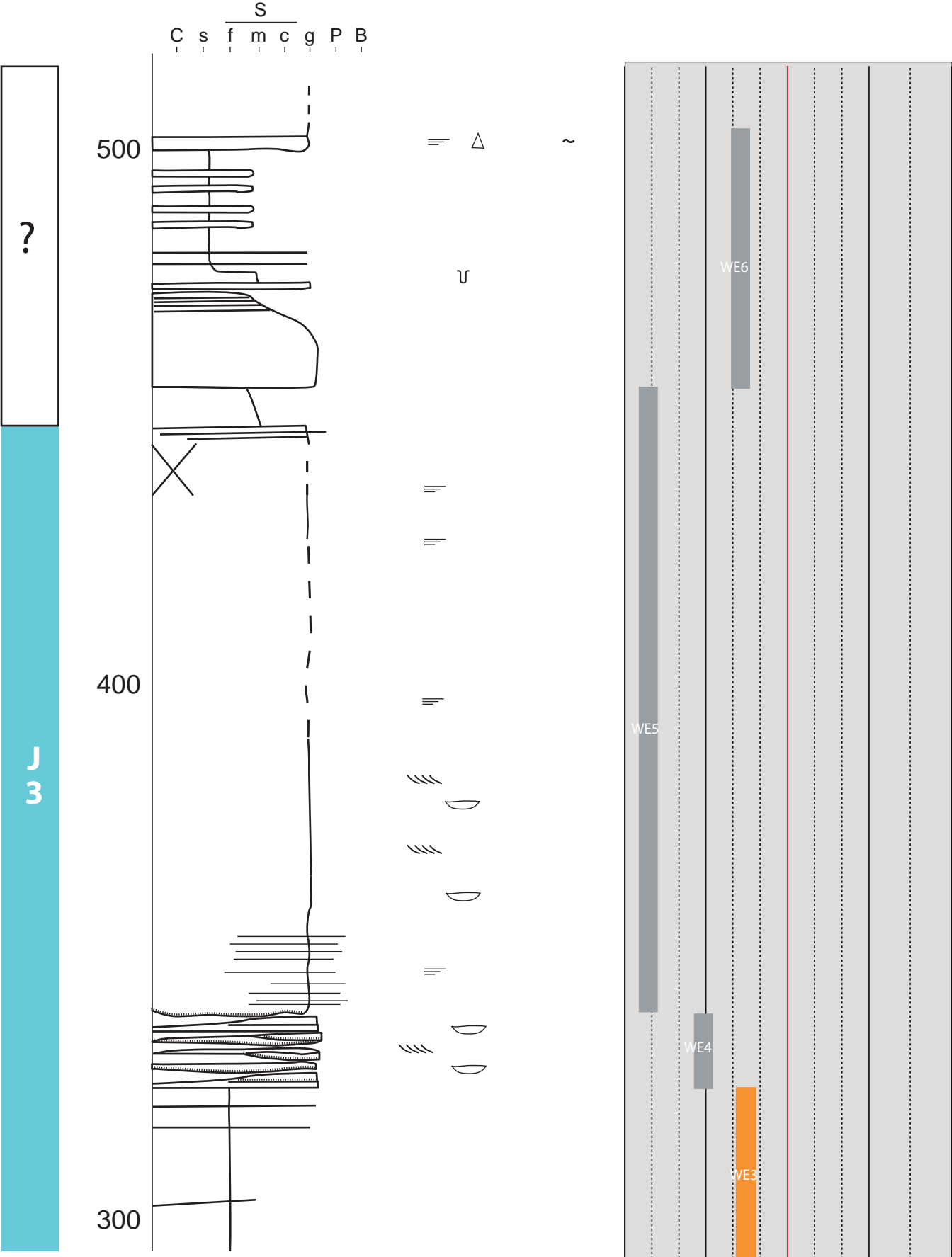


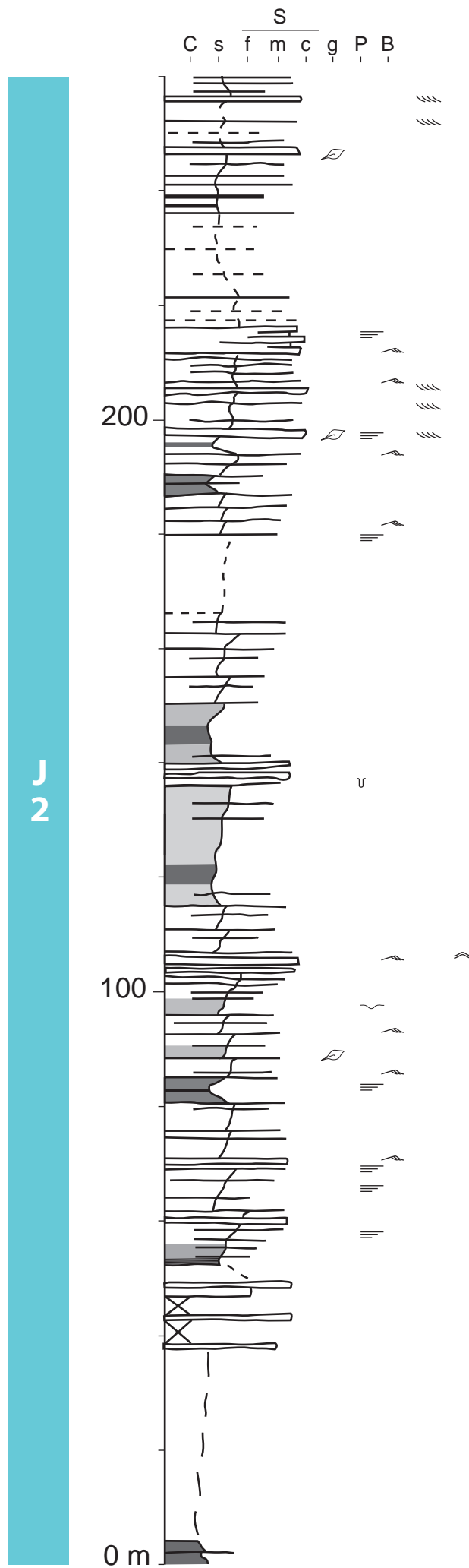
GPS W49

λ?

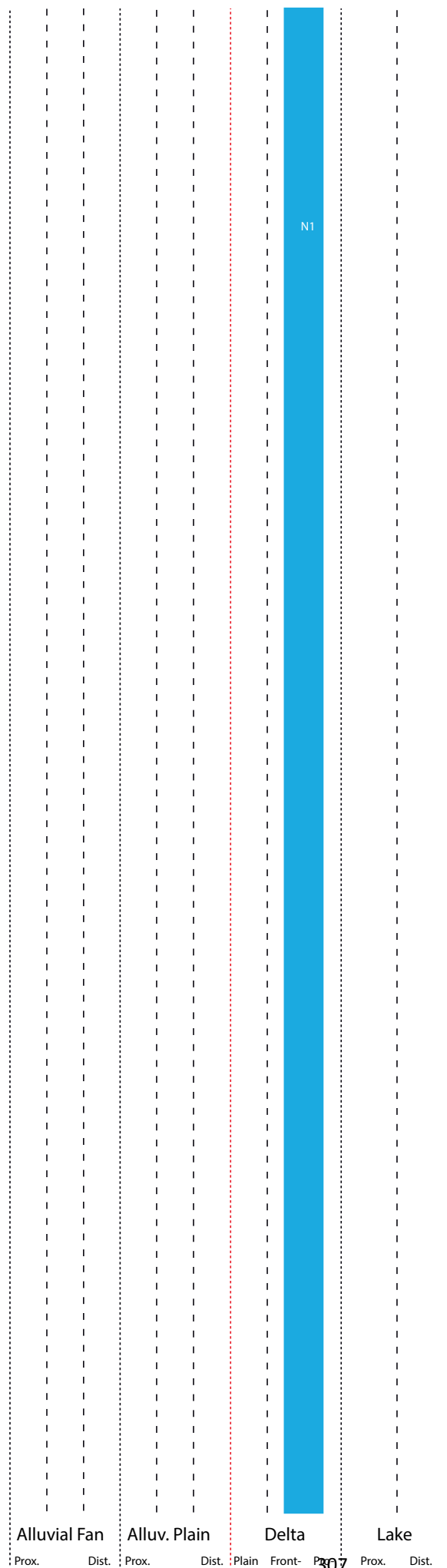




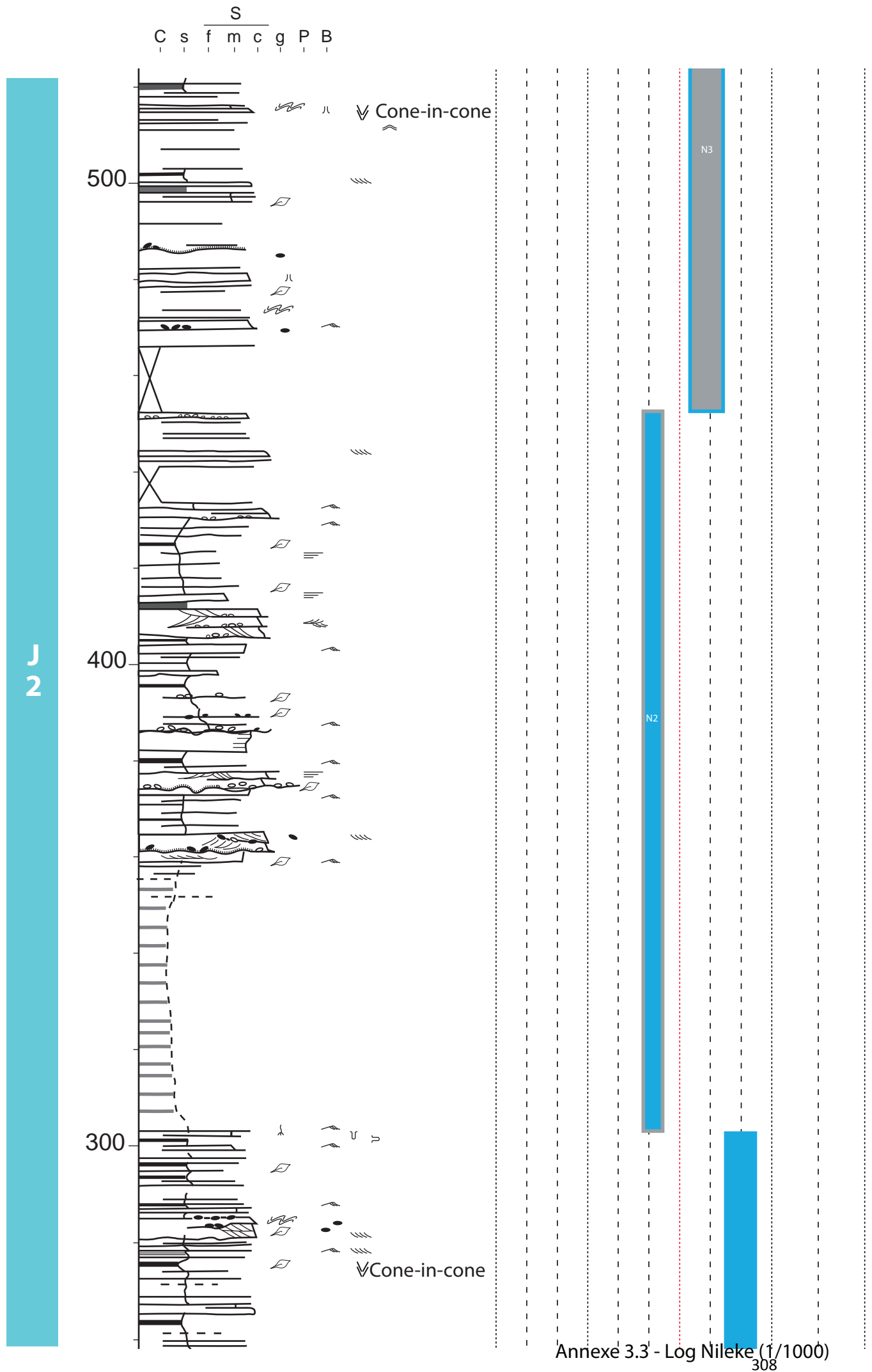


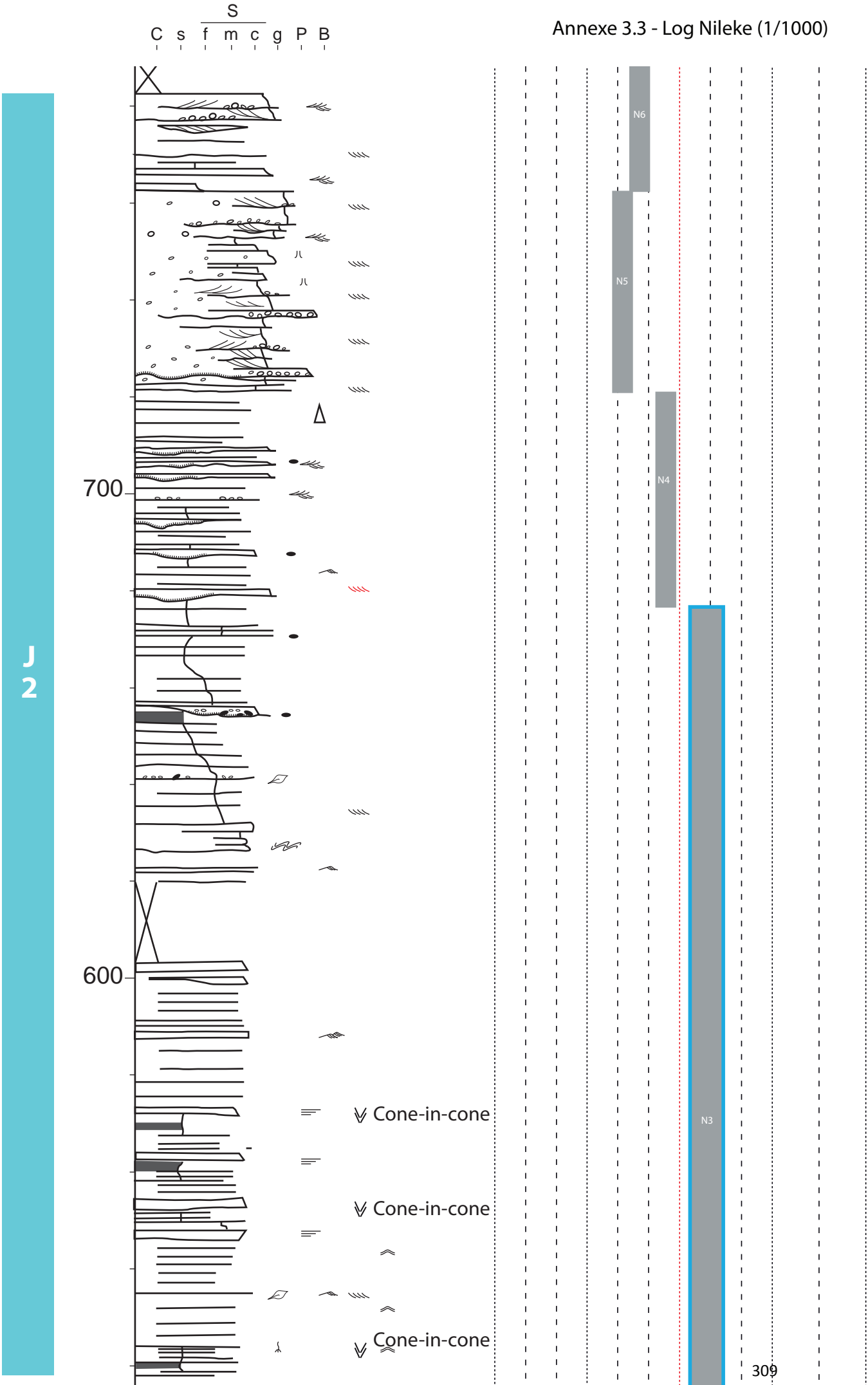


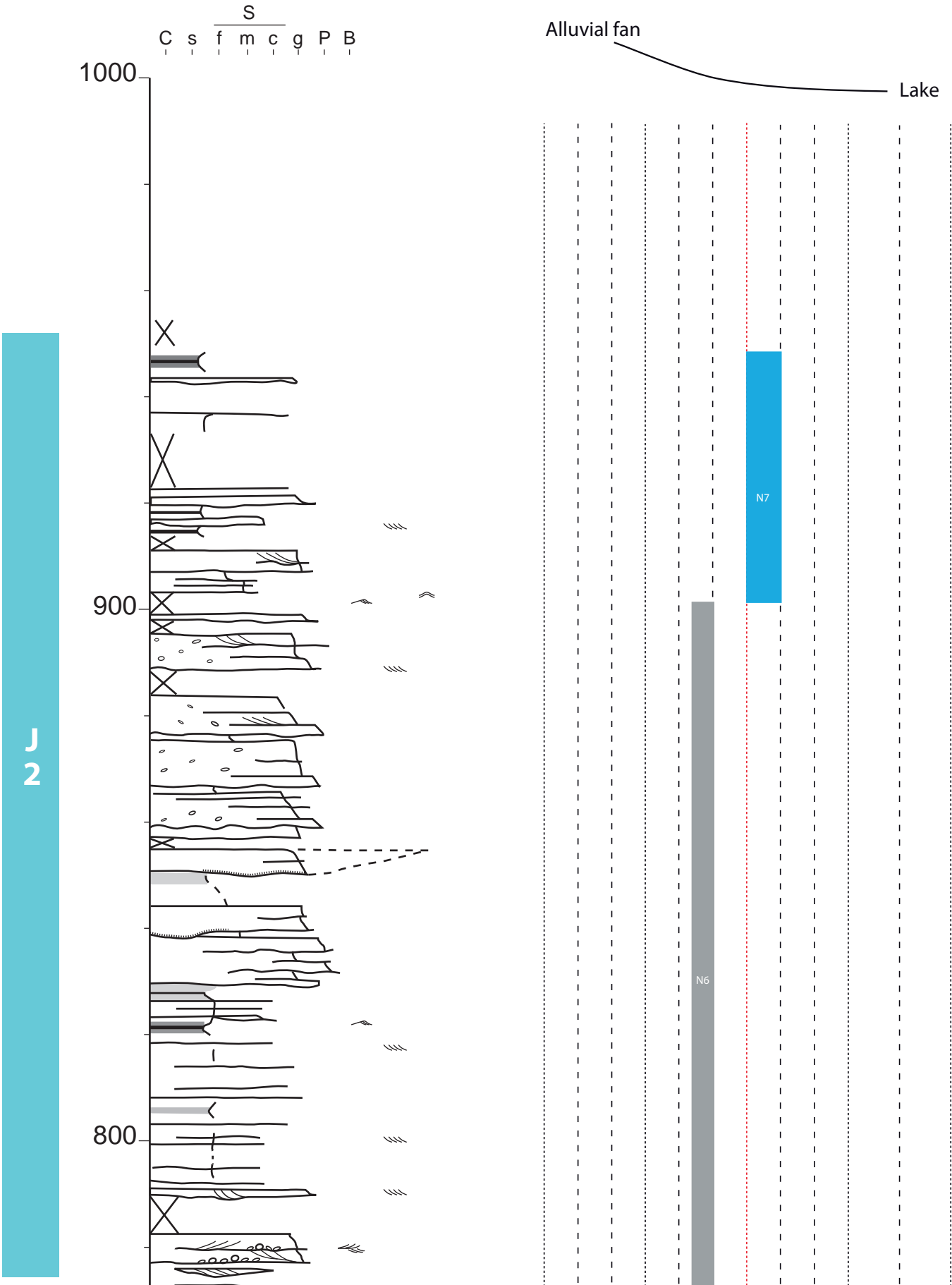
Annexe 3.3 - Log Nileke (1/1000)

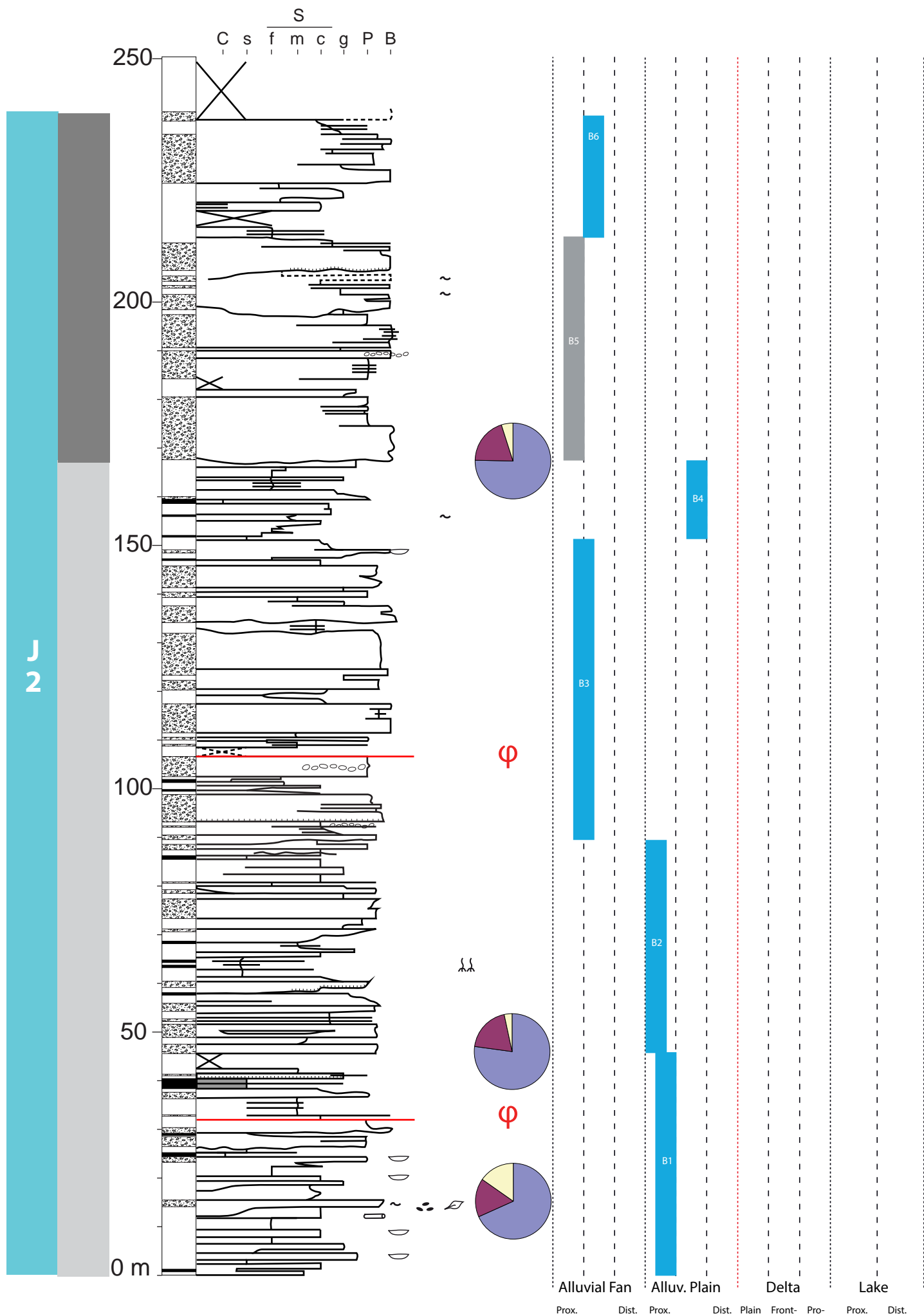


N1



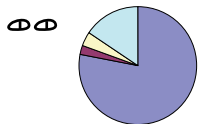
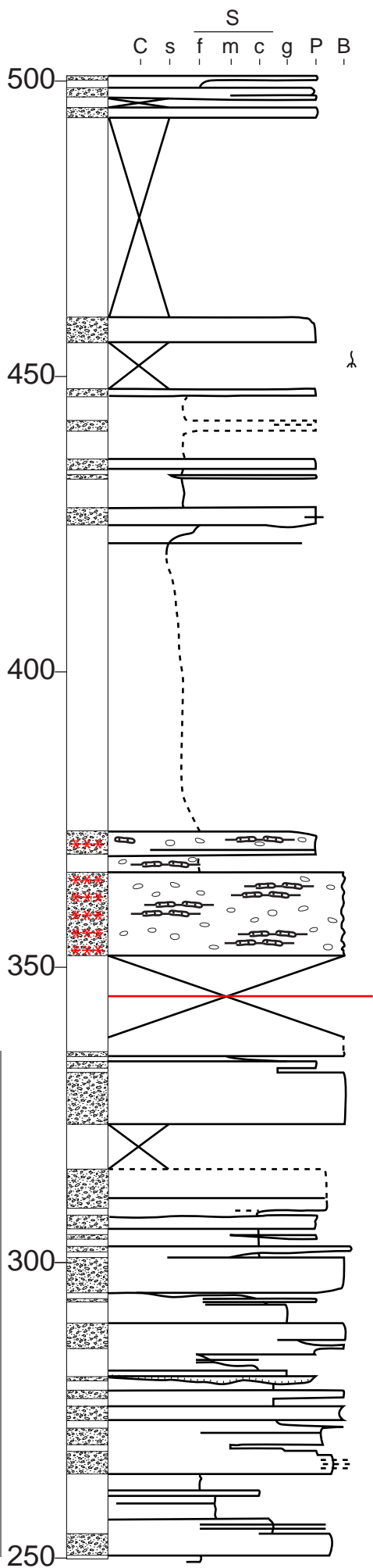






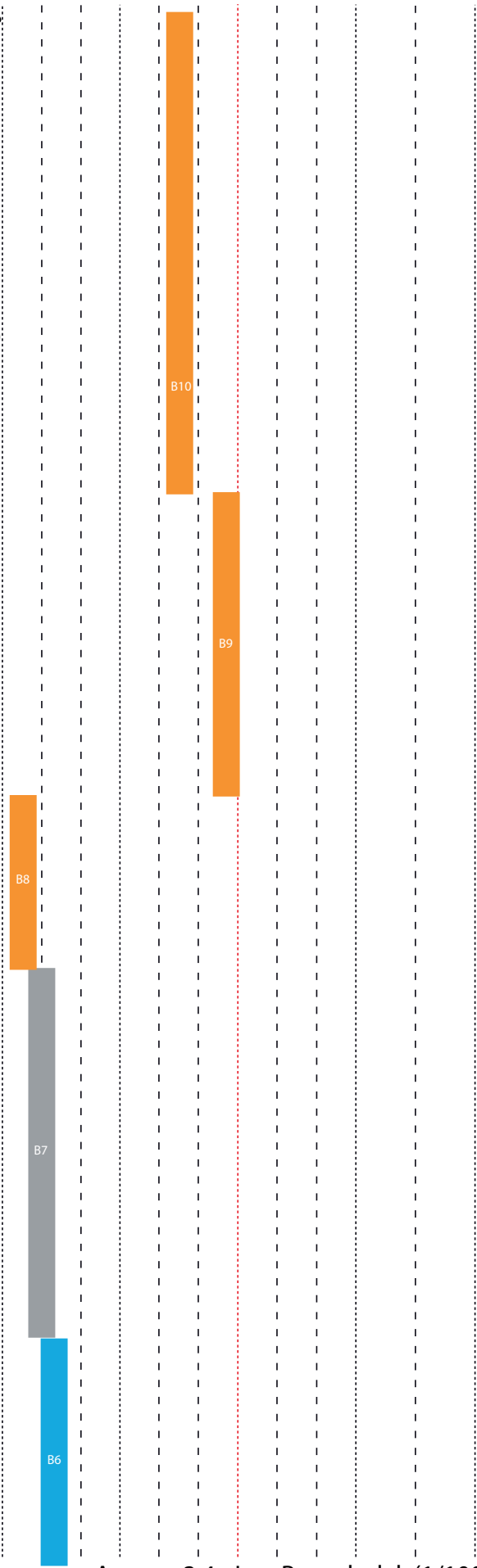
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Paleogene?

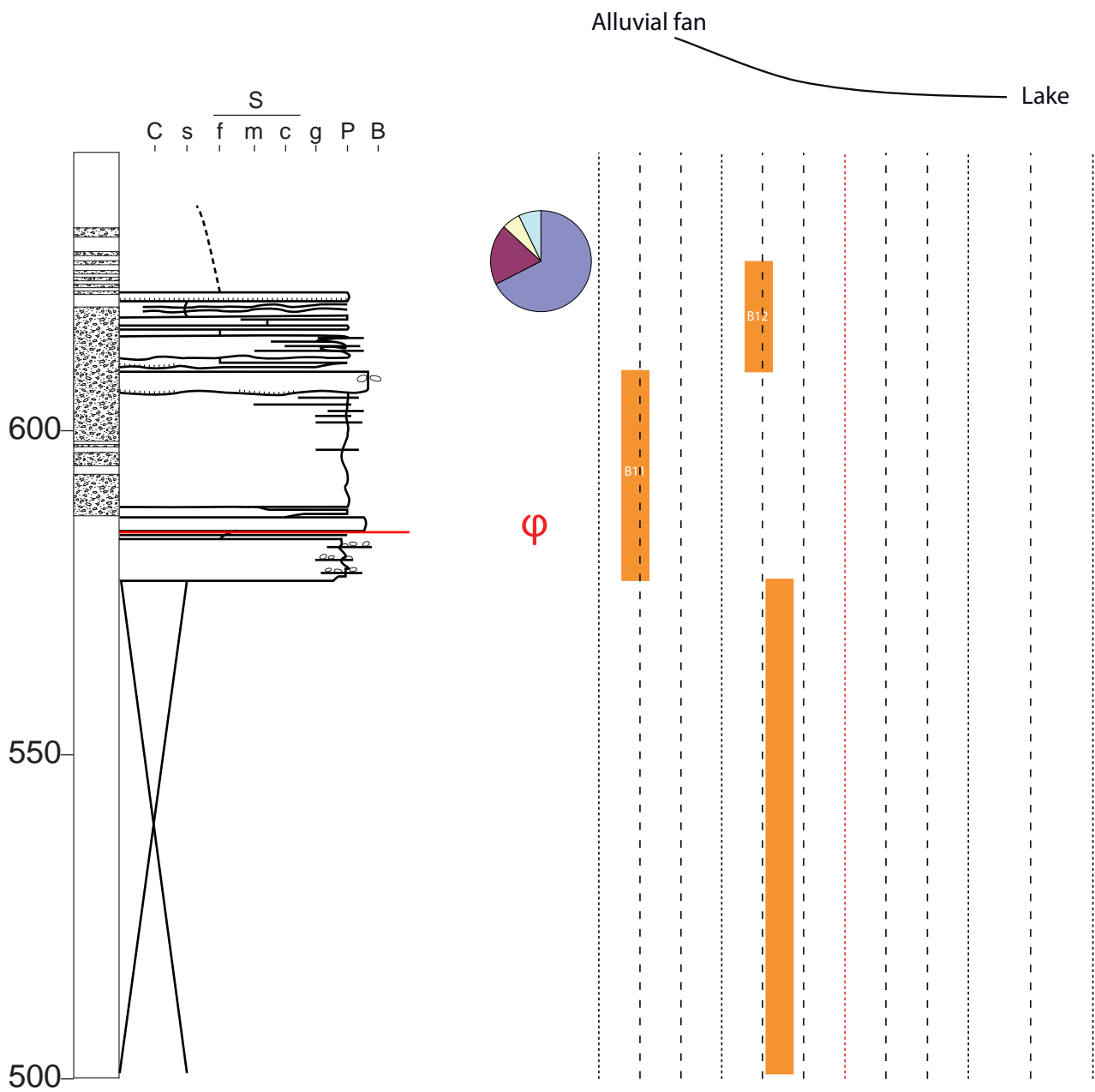


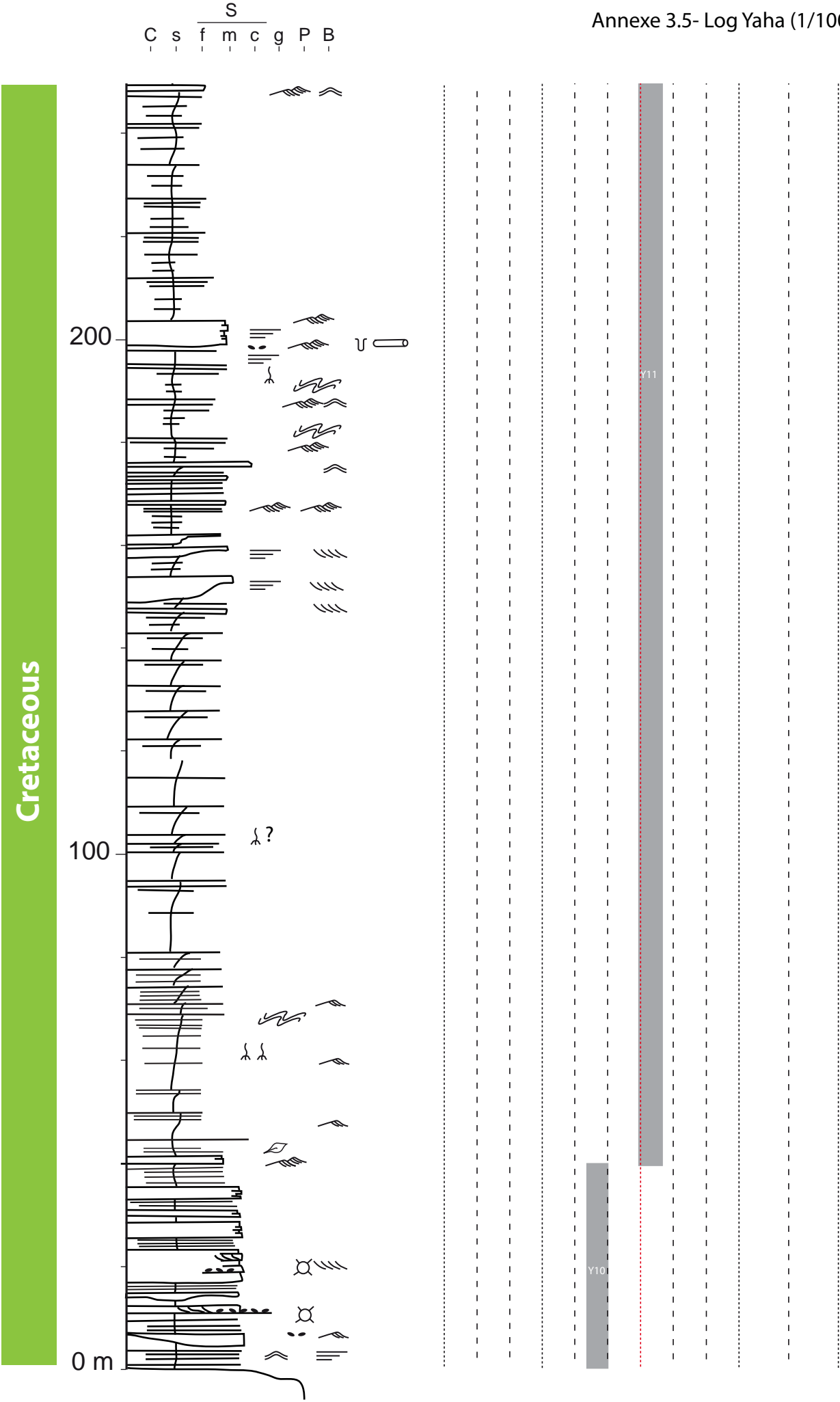
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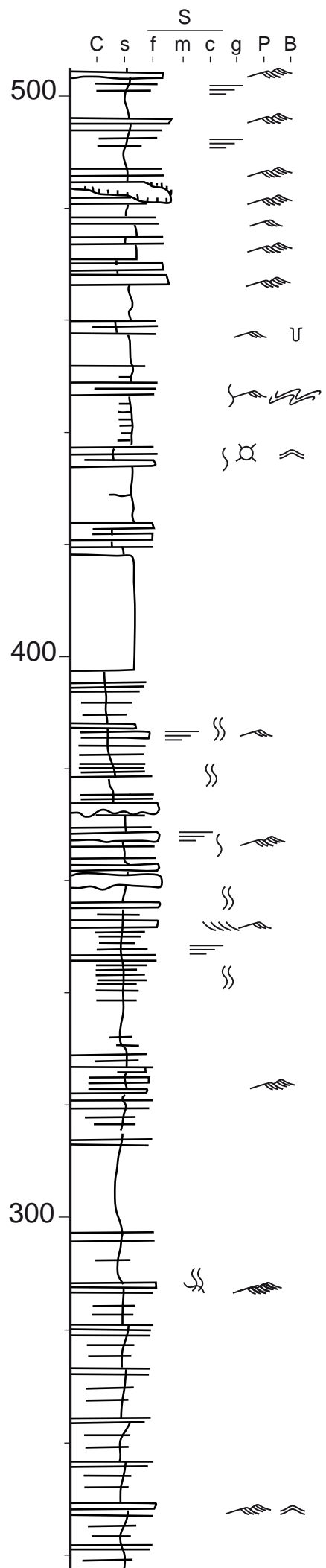


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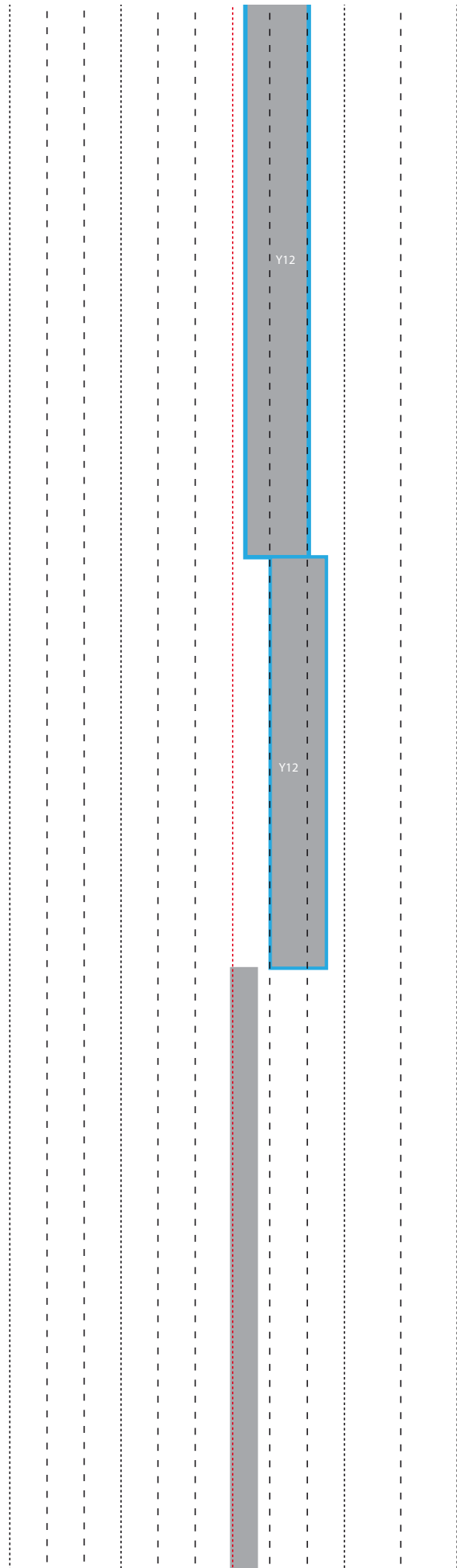


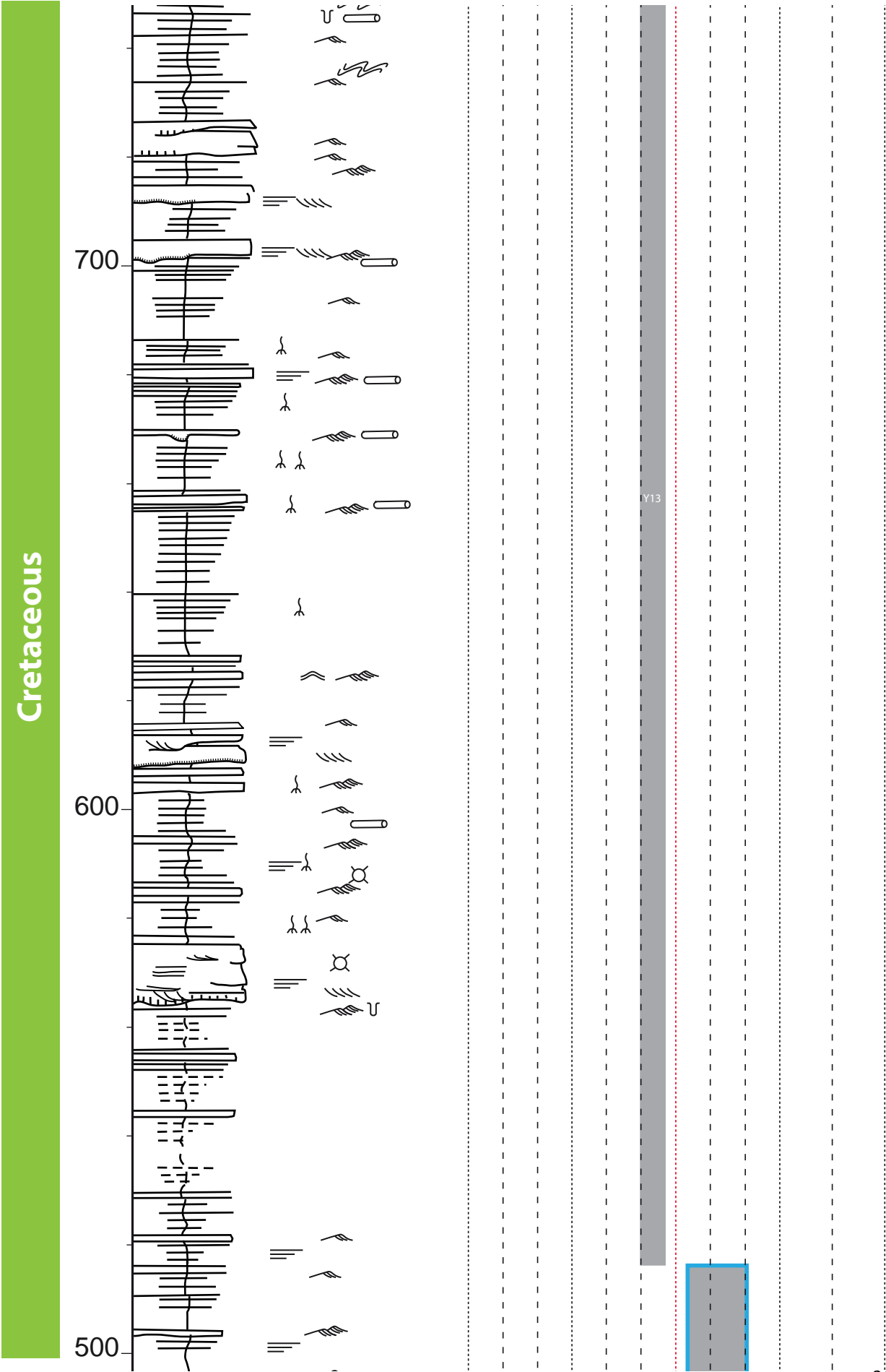


Cretaceous

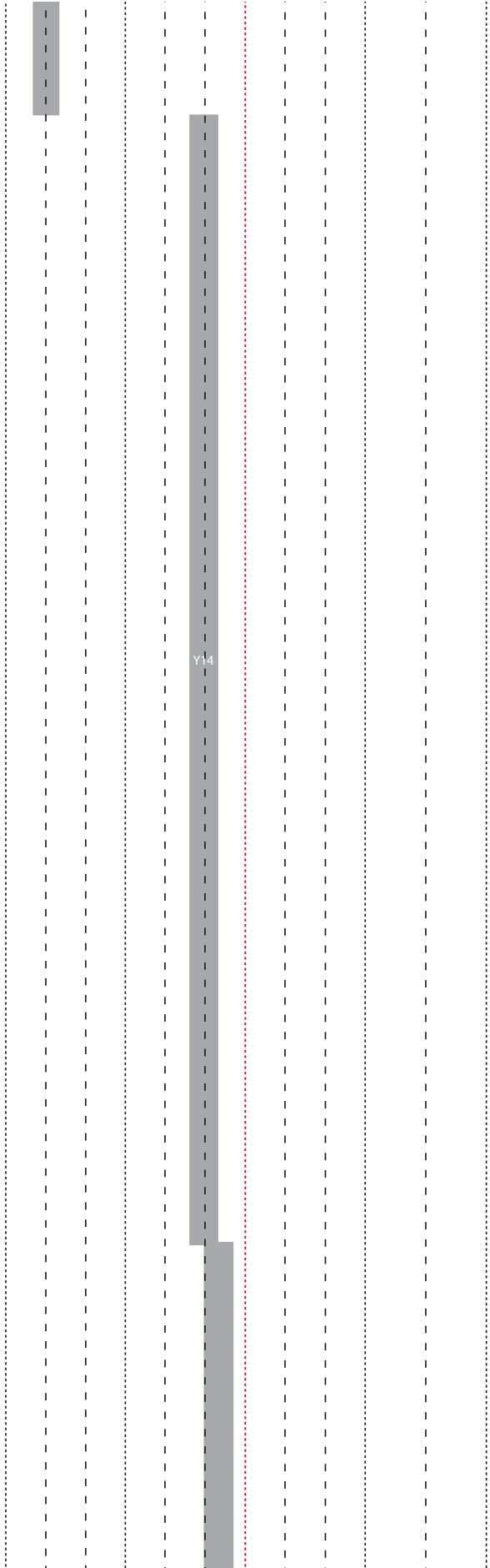
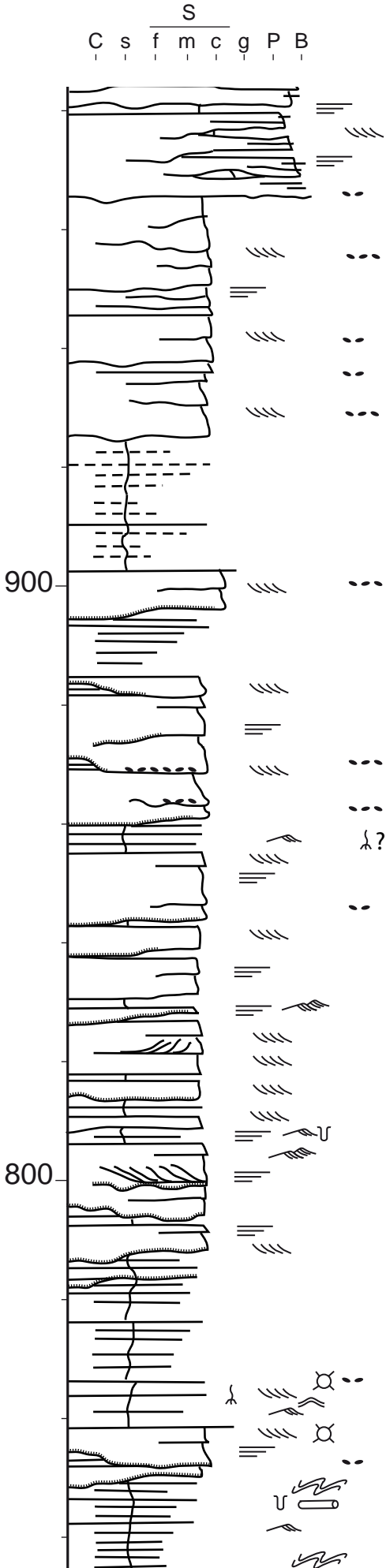


Annexe 3.5- Log Yaha (1/1000)





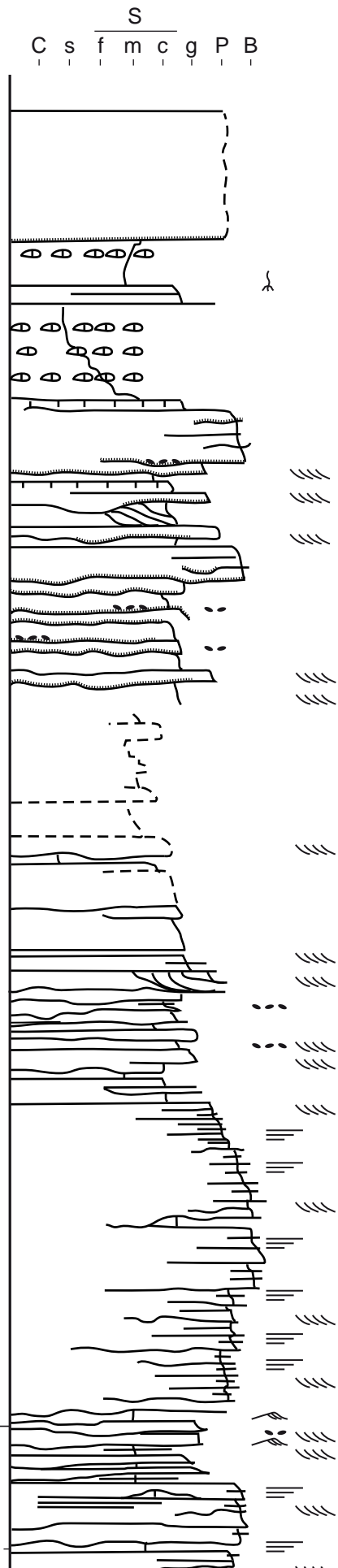
Lower Cretaceous



Paleog.

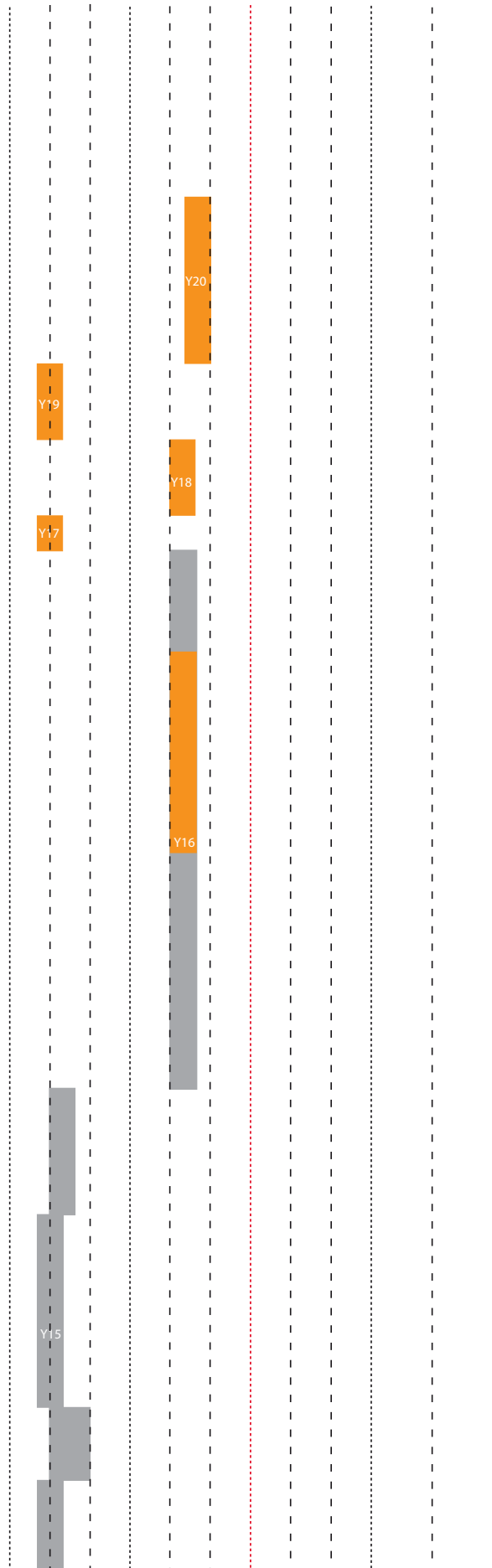
Upper Cretaceous

?



Alluvial fan

Lake



Annexe 3.5- Log Yaha (1/1000)